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METAMORPHISM IN THE CANADIAN SHIELD

1978

NOTE

Metamorphic Map of the Canadian Shield

In this volume frequent reference is made to Map 1475A "Metamorphic map of the Canadian Shield". This multicoloured map (scale 1:3 500 000) is available separately from all sales offices of the Geological Survey of Canada. Price $2.00 per copy.

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METAMORPHISM IN THE CANADIAN SHIELD

Proceedings of a Symposium held in Ottawa, Canada, May 5 - 6, 1977

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Foreword

Geological mapping of the Canadian Shield had reached the stage where it was possible by the early seventies to provide the first compilation and an overview of the history of metamorphism in the Shield. By 1974, partly in response to an international decision to produce a metamorphic map of North America and of the World, it was also decided to initiate a project to produce the first metamorphic map of the Canadian Shield. Because of the complexity of Shield metamorphism, it was necessary to obtain the help of many geologists who were brought together at a working symposium held in Ottawa, May 5-6, 1977. Twenty-three papers describing both local and regional problems were presented to one hundred delegates and a preliminary draft of the metamorphic map of the Shield on a scale of 1:1 000 000 was displayed for discussion and comment. This symposium volume, "Metamorphism in the Canadian Shield", is an outgrowth of the working conference.

The symposium volume and complimentary metamorphic map are expected to focus the attention of petrologists on metamorphic problems of the Canadian Shield and to result in many years of fruitful research.

It was particularly fitting that J.A. Fraser and W.W. Heywood should have assumed the overall co-ordination of this project. They were among the small band of leaders during the great era of helicopter-supported mapping that covered vast tracts of previously unknown territory, and that fundamentally altered our knowledge of the geological framework of much of the Canadian Shield.

W.F. Fahrig
Head, Precambrian Subdivision
Geological Survey of Canada

Ottawa, August 1978

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SOUTHERN PROVINCE

GRENVILLE PROVINCE

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A limited number of separates of the papers that appear in this volume are available by direct request to the individual authors. Authors' addresses appear on the last page of each paper.

Representation of metamorphism in granitic terranes is similarly hampered by the problem of recognizing metamorphic grade in granitic rocks. Metamorphism in some granite plutons has been inferred from the effects of metamorphism in spatially associated younger supracrustal rocks and isograds in the younger rocks have been projected into the granite. This procedure, which unfortunately conceals the original identity of the pluton, is commonly based on assumptions that cannot be verified.

Some of these problems may be resolved in the future. A program for data storage of metamorphic mineral assemblages currently being undertaken by the Geological Survey of Canada should ensure greater flexibility and precision in the representation of metamorphic data.

This volume and the Metamorphic Map of the Canadian Shield (Map 1475A) embody the efforts of many contributors. The names of the principal contributors, including authors of papers and compilers of regional data for the map, are shown in Figure 1 according to area of responsibility. It is a pleasure to acknowledge the enthusiasm and co-operation displayed by these participants in all phases of the metamorphic project. We wish also to thank W.S. Fyfe of the University of Western Ontario, who prepared the first paper in this volume as an opening address for the symposium, but was prevented from delivering it by sudden illness. Suites of thin sections from the Grenville Province and from Boothia Peninsula were made available through the generosity of A. Baer of the University of Ottawa and R.L. Brown of Carleton University. T.M. Gordon, A.N. LeCheminant, and A.V. Okulitch, all of the Geological Survey of Canada, supplied unpublished data for the map from the District of Keewatin. E. Froese of the Geological Survey of Canada arranged for submissions from geologists in Alberta, Saskatchewan, and Manitoba, and also acted as consultant on theoretical aspects of metamorphism. J. Bourne of the Geological Survey of Canada co-ordinated submissions from geologists working in the Grenville Province, and critically read the manuscripts. R. Skinner of the Geological Survey of Canada arranged for submissions from geologists in the Ontario Division of Mines. E. Froese, W.L. Davison, R.K. Herd, and G.D. Jackson, all of the Geological Survey of Canada, served with W.W. Heywood and J.A. Fraser on the committee that designed the map compilation scheme. Special thanks are due to Marcia A. Mazurski who not only assumed major responsibility in the day-to-day compilation operations but also, as metamorphic project petrologist, examined almost two thousand thin sections from northern Shield collections. We are indebted to P. Chernis for his able assistance in many facets of the work, including compilation and petrography. Finally, we thank B. Cox for typing many of the manuscripts, K. Bencik for drafting many of the figures, and W.F. Fahrig, Head of the Precambrian Subdivision, for his untiring support.

J.A. Fraser and W.W. Heywood
Co-ordinators, Metamorphic Map Project

INTRODUCTION

This volume consists of papers presented to the symposium 'Metamorphism in the Canadian Shield', which was held in Ottawa, May 5-6, 1977, to mark the completion of the first metamorphic map of the Canadian Shield (Map 1475A). The papers provided the regional contributors to the map project an opportunity to discuss the outstanding metamorphic features in their regions and also to place on record information too detailed to be shown on the map. Papers have been grouped under structural provinces which appear in approximate order of decreasing age; within each province and between adjacent provinces they have been arranged in an order that reflects geographic continuity.

The map was compiled at a scale of 1:3.5 million from about 30 regional maps prepared at a scale of 1:1 million by geologists having access to field records and thin section collections covering relatively large areas of the Shield. In many cases it was possible to use the data from these regional compilations as submitted; in other cases, however, discontinuities in unit boundaries between adjacent map areas, resulting from differences in levels of available information and differences in interpretation, required smoothing of boundaries and, in some instances, redefinition of metamorphic units. In the few places where differences were too great to be reconciled, original boundaries were left intact. The compilers accept full responsibility for the use made of the data received.

In view of the large number of contributors to the project it was essential that a standard system for compiling regional data should be followed. The metamorphic map of Europe, completed under the guidance of H.J. Zwart, Chairman for the Working Group of Metamorphic Belts of the World, provided a useful model of a system employing many metamorphic variables. The scheme adopted for the Metamorphic Map of the Canadian Shield was designed to accommodate the special characteristics of Shield geology as well as wide variations in data base. Modifications were made to the scheme where necessary as the compilation progressed.

The map shows the distribution of the five principal metamorphic facies as defined by Winkler in 1968 and 1974: subgreenschist, greenschist, lower amphibolite, upper amphibolite, and granulite. No distinction has been made between regional and contact metamorphism. Facies have not been divided into subfacies, but isograds, some of which coincide with boundaries of subfacies, have been drawn where data allow. Provision was made for portraying undivided facies ranges and for representing facies produced during two successive metamorphic episodes. Some estimate of pressures prevailing in specific regions can be inferred from isograd sequences in combination with the distribution of pressure indicator minerals such as andalusite, kyanite, and sillimanite. Occurrences of several other minerals useful in defining or delimiting facies and subfacies have been plotted, although the stability fields of some of these are as yet poorly known. Orthopyroxene, diagnostic of the granulite facies, has been shown only as isolated occurrences in terranes of lower grade. The grouping of two or more minerals at one locality does not imply mutual stability. The absence of any or all mineral symbols from particular regions may merely reflect a lack of information.

The distribution of unmetamorphosed intrusive rocks, broadly classified as to lithology, has been represented on the metamorphic map as well as the distribution of unmetamorphosed undifferentiated cover rocks. Lithologies of the metamorphosed rocks, on the other hand, are not indicated except in the Grenville Province, where there exists fundamental disagreement among the principal contributors as to the origin, whether magmatic or metamorphic, or granulite, charnockite, and anorthosite. Approximate ages of the unmetamorphosed rocks are designated by era or era ranges. However, the ages for the metamorphic rocks, similarly expressed, refer only to the ages of metamorphism.

Certain difficulties were encountered in using the compilation scheme. Perhaps the most serious limitation in delineating metamorphic facies boundaries stems from the lack, in many rocks, of diagnostic mineral assemblages. This is especially apparent in rocks metamorphosed at very low grade. Consequently, more rocks have probably been metamorphosed to subgreenschist facies than indicated. The widespread development of minerals such as muscovite, chlorite, and epidote, characteristic of metamorphism at low and very low metamorphic grades, has been described by several contributors. These minerals were not plotted, partly because of the problem of distinguishing low grade regional metamorphic effects from those associated with fault movement and/or produced by deuteric alteration, and partly because portrayal of a subgreenschist facies overprint of great areal extent would mask the distribution of the facies of higher grade. In many regions such representation is pre-empted by the necessity of depicting two previous metamorphic episodes. Boundaries between greenschist and amphibolite facies terranes in regions void of pelitic assemblages in some cases have been approximated using composition of plagioclase. In high grade terranes the association hornblende-sphene has locally served to distinguish rocks in the amphibolite facies from those in the granulite facies, and in the absence of more precise criteria, the presence of migmatite has been used to define the upper amphibolite facies. It is apparent from the map, however, that throughout much of the Shield vast amphibolite facies terranes could not be subdivided.

120°
80°
40°
Figure 1
G.D. Jackson
J.H. Maley
J.C. McGlynn
P.A. Nielsen
G.D. Jackson
W.C. Morgan
P.H. Thompson
R.A. Frith
W. W. Heywood
M. Schau
J.A. Fraser
60°
60°
K.E. Eade
A.N. LeCheminant
L. Westra
K. Schimann
J.F. Lewry
C. Rees
T.I.I. Sibbald
A.H. Bailes
W.D. McRitchie
D.C.P. Schledewitz
R.F.J. Scoates
W. Weber
R.K. Herd
F.C. Taylor
J.D. Godfrey
C.W. Langenberg
G.D. Jackson
I. Ermanovics
E. Froese
E. Dimroth
B. Dressler
J.H. Maley
J. Bourne
R. A. Burwash
(subsurface)
P.C. Thurston
F.W. Breaks
L.D. Ayres
J. Pirie
W.O. Mackasey
T. Clark
A. Franconi
E. Chown
R.K. Herd
J. Bourne
A.F. Laurin
J. Bourne
M. Rive
W.T. Jolly
J. Rondot
K.N.M. Sharma
G.B. Morey
M. Walton
K.D. Card
D.M. Carmichael
P.R. Whitney
100°
80°
60°

CRUSTAL EVOLUTION AND METAMORPHIC PETROLOGY

W.S. Fyfe[1]

Fyfe, W.S., Crustal Evolution and Metamorphic Petrology; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 1-3, 1978.

Abstract

Patterns in ancient metamorphic and igneous rocks are in accord with models of an early crust that was both thinner and hotter than the present crust. These rocks show no evidence for subduction or ocean-floor spreading processes. Regions with high thermal gradients appear to have been more closely spaced in the early crust. It is suggested that crust creation by "hot spot" phenomena may have been dominant.

Résumé

Les configurations visibles des roches anciennes métamorphiques et ignées correspondent aux modèles d'une écorce terrestre originale qui était à la fois plus chaude et plus mince que l'écorce actuelle. Ces roches ne témoignent d'aucun processus moderne de subduction ou d'expansion du fond océanique. Les régions caractérisées par des gradients thermiques élevés semblent avoir été espacées à intervalles plus étroits dans l'écorce terrestre première. On suggère que la création de l'écorce par des phénomènes de "points chauds" pourrait avoir été dominante aux époques anciennes.

INTRODUCTION

I was asked by the convenors of this conference to provide some opening remarks and I would like to comment briefly on what I think are some of the major problems we face in contemplating metamorphic patterns in ancient rocks.

A noted geophysicist once remarked that as plate boundaries are defined by seismic events therefore plate tectonics is basically a part of seismological science. While there is truth in this statement, there is also little doubt that the task of unravelling past tectonic regimes is very much the field of the petrologist.

Earth dynamics and crustal evolution are largely the result of convective heat-mass transfer from within the earth. While the present regime can be rather precisely documented by geophysical methods, we must depend almost solely on petrological approaches to understand the past.

THE MODERN REGIME

The modern convective regime is associated with a number of well-defined petrological events. Where convective cells rise from the mantle there is massive intrusion and eruption of mantle-derived basalt. In general, if this process is continuous, it becomes submarine. Following the igneous event, the cooling process involves large-scale seawater convection, chemical alteration, and hydrothermal metamorphism, resulting in the production of spilite. The spilite may be in the zeolite, greenschist or amphibolite facies of metamorphism. The ridge environment is one in which thermal gradients are high and laterally variable, and in which the metamorphic facies may be compressed into a thickness of a few kilometres.

Where subduction occurs deep fusion produces andesitic magmas. At present this usually occurs near or under continental crust. The accumulation of andesitic volcanic rocks thickens the crust, which results in melting of the crustal basement and the production of granitic rocks. Crustal stability appears to be rather closely controlled by conditions of partial fusion. An over-thickened crust develops a high-grade refractory root in the granulite or amphibolite facies, and an upper zone of rather high-temperature regional and contact metamorphism associated with rise of magmas.

Near the trench regions where subduction is initiated, the descent of cold lithosphere results in development of blueschists, which are characteristically produced at low thermal gradients. Large scale overthrusts and nappe formation are typical. Such structures are often easily recognized by inverted metamorphic gradients, which should in principle be transient, but in reality may be at least partially preserved if fluids cannot pervasively penetrate the overthrust section of a thrust.

Ocean-continent or continent-continent collisions are frequently recognized because the sutures become decorated with spilitic (ophiolitic) assemblages and in places by blueschists (e.g. the Urals).

Where modern plate motion is accompanied by transcurrent motion, as along the Alpine Fault of New Zealand and the San Andreas Fault of California, large shear zones are developed. Such faults appear to be related to very thick sections of trench sediment debris, and perhaps a necessary condition for their operation is the presence of rapidly accumulated cold and wet sediment piles. As these return to a more normal thermal gradient they periodically release (the seismic pumping mechanism) very large volumes of fluids along the fault planes, which are commonly heavily mineralized (Sibson, et al., 1975).

In general, were all mantle motions to cease, the petrologist would have little difficulty interpreting the thermal and hence, convective pattern that led to the present crust.

The recent developments in global tectonics have many important consequences in terms of our thinking about the past. Some that I would stress are:

(a) The present rate of overturn of the crust is such as to modify about 80 per cent of the surface in 200 million years. In the past, the rate was almost certainly much faster (Elder, 1976, p. 83).

(b) Understanding heat flow regimes is one of the principal clues to understanding tectonic style; heat flow patterns are well recorded in metamorphic rocks.

(c) Major tectonic events may have occurred dominantly in a submarine environment; they may result in a deep circulation of water that could drastically modify the chemistry of the rock and virtually control the thermal structure. Thus the cover rocks of an igneous intrusion may finally crystallize as low-temperature, zeolite facies rocks and not contact hornfels rocks. However, the

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interplay of such factors can be resolved by the use of many features such as oxygen and hydrogen isotopes, oxidation-reduction phenomena and metasomatic events.

(d) Most of the major regional prograde metamorphic events occur where the crust is thickened by volcanic-tectonic processes. Commonly the overthickened part of the crust is removed by erosion and the fragmentary metamorphic events recorded in the root zone reflect such crustal disequilibrium.

CALIBRATION OF THE METAMORPHIC RECORD

One of the great advances in metamorphic petrology of the past three decades has involved the laboratory calibration of many metamorphic reactions. In many cases simple reactions can provide significant information. Thus the blueschist facies gradient of $10^\circ\ km^{-1}$ or less, is required by simple reactions such as:

calcite $\longrightarrow$ aragonite

albite $\longrightarrow$ jadeite + quartz

In other regimes a wealth of information can be derived from Al_2SiO_5 polymorphism plus simple reactions such as those involved in muscovite breakdown, but the use of such data requires careful consideration of kinetic phenomena. A few polymorphic reactions may be reversible; some ancient kyanite, for example, is a replacement of earlier andalusite (Fyfe, 1976).

Element distribution reaction, particularly in higher grade rocks, is becoming a popular method for calibrating pressures and temperatures. However, great caution must be applied where experimental data are sparse and thermodynamic extrapolations are used. There are many possible thermodynamic models of a solid solution and there is a great temptation to use the model that gives the "reasonable" result. Increasingly, we use $^{18}O/^{16}O$ distributions to estimate temperatures, although commonly the calibrations leave much to be desired.

All such mineral memory units must have a limit controlled by diffusion. Above a certain temperaure range the phases will continuously readjust to a new temperature. The situation is analagous to the limitations of simple potassium-argon dating.

In general, we have few data on the relevant diffusion coefficients but for this particular problem new techniques are available which could improve the situation. With modern ESCA and Auger spectroscopy, we can analyze the outermost 1 to 5 nm of a solid for most elements. If a crystal is chemically zoned, we can observe changes in concentrations in a very thin zone and obtain diffusion parameters even when the diffusion process is extremely slow. Using the diffusion relation:

$$\bar{x}^2 = 2Dt$$

where $\bar{x}$ is the mean diffusion distance, t the time and D the diffusion coefficient, it will be possible to measure diffusion over distances of 10 nm or so even when D is in the order of 10^{-20}. Abundant new data could be obtained from such studies. Most such measurements have previously been restricted to igneous, rather than metamorphic, temperature ranges. In addition, these thin-film techniques could add to our knowledge of the distribution of certain minor elements in a mineral assemblage, and indicate whether or not they are concentrated on grain boundaries.

AGE DETERMINATION

At the heart of any tectonic synthesis is the problem of exact age determination. In any regime, we wish to know the age of the source rocks and the timing of every event leading to modification of this age. Every method of determining age has its limitations, including those imposed by diffusional reequilibration and problems of sampling. The truth is, that for a metamorphic problem we need some combination of all the methods, from those that date whole rocks to those that date separate minerals. Each will record a useful piece of information. The recent work of Boger and Faure (1976) on the apparent Rb-Sr age of 230 million years for Red Sea sediments nicely illustrates the sampling problem. What do such ages mean in an "ancient gneiss" complex?

I think that when the ion-microprobe reaches its full potential, we will see great advances resulting from zircon dating, particularly of zoned zircon crystals. And if eventually the ion-probe can be used to determine attributes such as $^{18}O/^{16}O$ zoning, we will be provided with much useful new information. Couple these techniques with more precise information on diffusion, and I am sure we will see spectacular advances in our appreciation of the metamorphic process.

THE ANCIENT CRUST

One of the most important problems of modern geology and in particular petrology, is to ascertain how long modern convective styles and processes have operated. A further question involves changes in the thermal gradient of the crust through time, and as a consequence, possible crustal thickness. In searching for answers to these questions, we are particularly concerned with the most ancient rocks, for if changes do occur it is in these that they should be most apparent. At present there is evidence to suggest that the ancient Earth was hotter than at present but convincing documentation is lacking. Perhaps the best evidence comes from hypotheses on the thermal evolution of the Moon and Mars.

Recently Hargraves (1976) presented a stimulating synthesis of crustal evolution. In essence, he proposed a very rapid differentiation of the Earth, the formation of a globe-encircling ocean, and a more or less continuous sialic (granitic) crust. The equilibrium crustal thickness is controlled by granite melting relations. Hargraves suggested that continental crust was reworked and thickened through time until eventually the continents emerged above sea level. Critical to this model is the concept of an almost constant mass of continental crust with a thickness controlled by the thermal regime. Not all agree with such a model, in particular Moorbath (1977), who proposed that sialic material was added to the crust over a long period of time. His arguments are largely based on $^{87}Sr/^{86}Sr$ data, and on the fact that most early "granitic" rocks have initial ratios essentially similar to the mantle. In general, geological maps support the Hargraves model and explain why crust-mantle mixing was rapid and has produced the effects which Moorbath observed.

But our problem is to find data which can reveal tectonic processes. Would the present style explain the nature of the ancient crust?

When we consider the question of thermal gradients certain features seem to be in accord with the hotter, thinner crust model. These might include:

(a) rarity of kyanite in ancient rocks (but beware of kinetic prograde and retrograde conversions)

(b) total lack of eclogites

(c) total lack of blueschists

(d) high temperature lava (komatiites)

(e) the apparent predominance of low-pressure granulites (C. Herzberg, pers. comm.).

When we consider ancient metamorphic phenomena, we must constantly bear in mind that regional metamorphism and partial melting may have resulted from short-lived overthickening as in the present.

When we consider tectonic styles additional differences are apparent. Ophiolite complexes of the modern type seem rare in ancient crust and acid igneous events appear to be concentrated in the submarine environment. Volcanic centres and zones of intense hydrothermal discharge (leading to ore deposits) appear to be closely spaced (hundreds, rather than thousands, of kilometres in the past). Low-grade metamorphic equivalents of ancient submarine basalts are common. However the closely compressed facies of the modern ocean floor are not common which suggests different crustal structures.

Kerrich (see Fryer, et al., 1977) has observed that if the crust was thinner, and confining pressures in thin crust lower, then deep convection of marine waters would be a more pervasive process. This effect could explain the more common development of large gold-quartz systems of the Yellowknife type and could, in part, explain $^{87}Sr/^{86}Sr$ homogenization by equilibration with early marine waters. Recent work by Lewis and Snydsman (1977) suggests that at present marine fluids may penetrate the ocean-floor crust to hydrate the mantle. Thus understanding fluid convective regimes in ancient crust is fundamental to understanding the significance of thermal patterns.

When we contemplate such observations we are left with some important questions. If there is no evidence for modern-type subduction metamorphic processes, what was the subduction process? Some form of subduction is surely required. Perhaps the entire convective process was different in that mantle convection occurred beneath a convecting crust.

The distribution of igneous bodies appears to leave little doubt that hot regions were much more closely spaced than at present. But was there no ocean-floor crust of the modern type and hence no major ocean-floor spreading process? In this connection I am impressed by Elder's (1976) work on thermal structures in a convecting body. One of the features he emphasized is that as the thermal gradient becomes steeper, and the system hotter, the radial temperature distribution shows greater fluctuations. Such basic phenomena could well explain the almost simultaneous eruption of normal basalts and komatiites.

Considering such observations, I think that we must keep an open mind on tectonic styles in the ancient crust. There is good evidence that the modern style and dimensions of convection do not necessarily apply to the Archean. While at present "hot spot" crust formation makes up a very small part (perhaps one per cent) of crustal creation, it may have been the dominant form of activity in a more turbulent, hotter earth. But the closely spaced volcanic "hot spots" would be penetrating, both over- and under-plating a thin granitic shell. The underplated material would cool slowly to generate anorthosites. The updoming associated with a rising plume, and the volcanic loading, could lead to granite production at depth (reworking), and lateral thrusting to cause short-lived overthickening. In such a model much of the subduction phenomena would involve simple vertical motion during plume decay, and the tectonics would be essentially of Ramberg (1967) type. The development of the craton would be associated with changes in the basic convectic pattern as was suggested by Runcorn (1965). With this model (cf. Hawaii on a granite basement) a thin granitic layer would be sandwiched between extrusive basalt on top and gabbro-anorthosite below. Considering the partial melting near the base, the prograde metamorphism at depth leading to rubidium loss, and the seawater convection through the upper layers, it is not surprising that the strontium record reflects mantle more than crust (Fryer, et al., 1977).

The flood of papers on Archean structures strongly suggests that we are a long way from adequately understanding them although we have seen great progress in recent years. Detailed mapping and careful petrologic studies will be required using every technique available to read the record of physical conditions and processes in old rocks which often reflect more than one metamorphic event. Most workers with whom I discuss such problems agree that the present style of tectonic processes differs from that of the Archean.

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ARCHEAN AND PROTEROZOIC METAMORPHISM IN THE NORTHWESTERN SUPERIOR PROVINCE AND ALONG THE CHURCHILL-SUPERIOR BOUNDARY, MANITOBA

W. Weber[1] and R.F.J. Scoates[1]

Weber, W. and Scoates, R.F.J., Archean and Proterozoic metamorphism in the northwestern Superior Province and along the Churchill-Superior boundary, Manitoba; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 5-16, 1978.

Abstract

A progressive increase in metamorphism from amphibolite facies to granulite facies in the High Hill Lake greenstone belt, at the south margin of the Pikwitonei region, indicates that the granulite facies metamorphism in the Pikwitonei region is younger than the deposition of the greenstone belt successions in the Cross Lake subprovince, and not older as previously suggested. Radiometric ages indicate that the rocks in the Cross Lake subprovince have undergone metamorphism between 2400 and 2800 Ma. Two metamorphic-plutonic events, at approximately 2700 and 2400 Ma, have been recognized in the Oxford Lake area and are inferred elsewhere.

Retrogression of Pikwitonei region granulites and refolding of Pikwitonei region structures along the Superior Province margin indicate that an amphibolite facies metamorphic-plutonic event overprinted Pikwitonei rocks in the Thompson Nickel Belt, the Orr Lake Segment, the Split Lake Block, and south of the Fox River Belt. On the basis of radiometric and structural data in the Churchill-Superior boundary region, this event is interpreted as the imprint of the Hudsonian front 1700-1800 Ma ago.

Low grade, essentially undeformed supracrustal rocks of the Fox River Belt occurring between intensely deformed high grade gneiss of the Superior and Churchill provinces appear to have been deposited after the main phase of the Hudsonian Orogeny. A Rb-Sr whole rock isochron age of 1600 Ma is interpreted as dating the major period of intrusion and volcanism in the Fox River Belt. The lithologically similar supracrustal Ospwagan group apparently occurs as infolded keels in Moak Lake gneiss near the western margin of the Thompson Nickel Belt. The Ospwagan group and Fox River Belt may be of similar age.

Rocks of the Churchill Province adjacent to the Superior Province consist of amphibolite facies, Aphebian metasedimentary gneiss. Distinct differences in lithology and age between Churchill Province and Superior Province rocks are used to define the boundary between the Churchill and Superior provinces, even in areas where the Hudsonian Orogeny apparently affected the rocks on both sides of the boundary.

Résumé

Une augmentation progressive du métamorphisme du faciès amphibolite au faciès granulite, dans la zone des roches vertes de High Hill Lake à la lisière sud de la région de Pikwitonei, indique que dans cette dernière région, le métamorphisme du faciès granulite est plus récent que les successions de la zone de roches vertes de la sous-province de Cross Lake, et moins ancienne qu'on ne le supposait auparavant. Les datations effectuées par des méthodes radiométriques indiquent que les roches de la sous-province de Cross Lake ont été soumises à un métamorphisme datant de 2 400 à 2 800 Ma. Deux évènements, de nature plutonique et métamorphique, se situant il y a environ 2 700 et 2 400 Ma, ont été identifiés dans la zone d'Oxford Lake, et l'on suppose qu'ils ont laissé leur marque.

Le métamorphisme régressif qu'ont subi les granulites de la région de Pikwitonei, et une seconde phase de plissement qui dans cette région a affecté la lisière de la province du lac Supérieur, montrent qu'un évènement plutonique et métamorphique d'intensité correspondant au faciès amphibolite, a aussi laissé son empreinte sur les roches de Pikwitonei, dans la zone de Thompson Nickel, le segment d'Orr Lake, le bloc de Split Lake et le sud de la zone de Fox River. D'après les données provenant d'études radiométriques et structurales de la bordure de la province de Churchill et de la province du lac Supérieur, on a déduit que cet évènement correspondait à la mise en place du front hudsonien, il y a 1 700 à 1 800 Ma.

Les roches supracrustales faiblement métamorphisées et très peu déformées de la zone de Fox River, situées dans un gneiss de haut degré métamorphique et déformé entre les provinces de Churchill et du lac Supérieur semblent s'être déposées après la phase principale de l'orogenèse de l'Hudsonien. L'âge de 1 600 Ma obtenu par la méthode des isochrones Rb-Sr appliquée à la roche entière, correspond sans doute à la date de la principale période d'intrusion et de volcanisme dans la zone de Fox River. Le groupe supracrustal d'Ospwagan, similaire du point de vue de la lithologie, forme apparemment des racines par fusionnement des plis, dans le gneiss de Moak Lake près de la limite ouest de la zone de Thompson Nickel. Il est possible que le groupe d'Ospwagan et la zone de Fox River aient le même âge.

Les roches de la province de Churchill, à proximité de la province du lac Supérieur, sont constituées d'un gneiss métasédimentaire d'âge aphébien, formé dans le faciès amphibolite. Les différences nettes de lithologie et d'âge entre les roches de la province de Churchill et celles de la province du lac Supérieur, servent à définir la limite entre ces provinces, même dans les zones où l'orogenèse de l'Hudsonien a apparemment modifié le caractère des roches de part et d'autre de cette limite.

[1] Manitoba Mineral Resources Division, Department of Mines, Resources and Environmental Management

REGIONAL SETTING

The region under discussion encompasses the Churchill-Superior boundary region and the northwestern Superior Structural Province, referred to as the Cross Lake subprovince by Stockwell et al. (1970).

Bell (1971a, b) subdivided the region into the "Wabowden subprovince", part of the Churchill Province, and the "Pikwitonei Province", which he separated from the Superior Province. However, the "Pikwitonei Province" and "Wabowden subprovince" are now considered to be part of the Superior Province and Bell's subdivisions are therefore unwarranted. This conclusion is based on the following observations:

a) The structural grain is continuous from the Superior Province into the "Pikwitonei Province". The granodiorite below the unconformity at Cross Lake and the granulite facies metamorphism are interpreted as Kenoran and not pre-Kenoran, as suggested by Bell (1971b). Thus, there is no evidence that the "Pikwitonei Province" is a basement to Superior greenstone belts.[1]

b) Rocks of Bell's "Pikwitonei Province" and "Wabowden subprovince" are lithologically similar and share much of their geological history, and in many places a broad transition occurs between the two domains. The "Wabowden subprovince" has been overprinted by the Hudsonian(?) Orogeny, and hence Bell (1971b) considered it as part of the Churchill Province. However, the fundamental geological contact between the Churchill and Superior provinces coincides with the western margin of the "Wabowden subprovince". This contact better satisfies the definition of structural province boundaries (Stockwell et al., 1970) than does the transition zone between the "Wabowden subprovince" and the "Pikwitonei Province".

Alternatively, we propose that the term "Pikwitonei region" be used in preference to "Pikwitonei Province" and that the term "Thompson Nickel Belt" (Coats et al., 1972) be used to encompass the rocks of the "Wabowden subprovince" (Fig. 1).

A linear belt of sedimentary, volcanic, and intrusive rocks, the Fox River Belt (Scoates, 1975a, b, 1977) characterized by lower metamorphic grade, less deformation, and different lithology than the adjacent rocks, is interpreted as overlying the Superior Province near its junction with the Churchill Province in the north. A similar suite of rocks, the Ospwagan group (Scoates et al., 1977), occurs in the Thompson Nickel Belt.

Highly metamorphosed metagreywacke and metashale strata of the Kisseynew Sedimentary Gneiss Belt (Bailes and McRitchie, 1978) are characteristic of the Churchill Province adjacent to the Thompson Nickel Belt. Similar metasedimentary gneiss forms the margin of the Churchill Province along the northern boundary of the Superior Province in Manitoba.

SUPERIOR PROVINCE

The northwestern part of the Cross Lake subprovince consists predominantly of layered felsic gneiss, amphibolite, and foliated to massive granitoid masses. Such rocks are poor indicators of metamorphic conditions with the exception of granulite facies. Thus, the Superior greenstone belts to the south (Fig. 1) provide the main data source for an analysis of the metamorphic history. East, northeast, and southeast trending faults cross-cut the Superior Province (Fig. 1). Some of these represent zones of vertical displacement suggesting that rocks from various crustal levels are exposed. Haugh (1969) suggested that the Split Lake Block (Fig. 1) may represent an uplifted crustal segment.

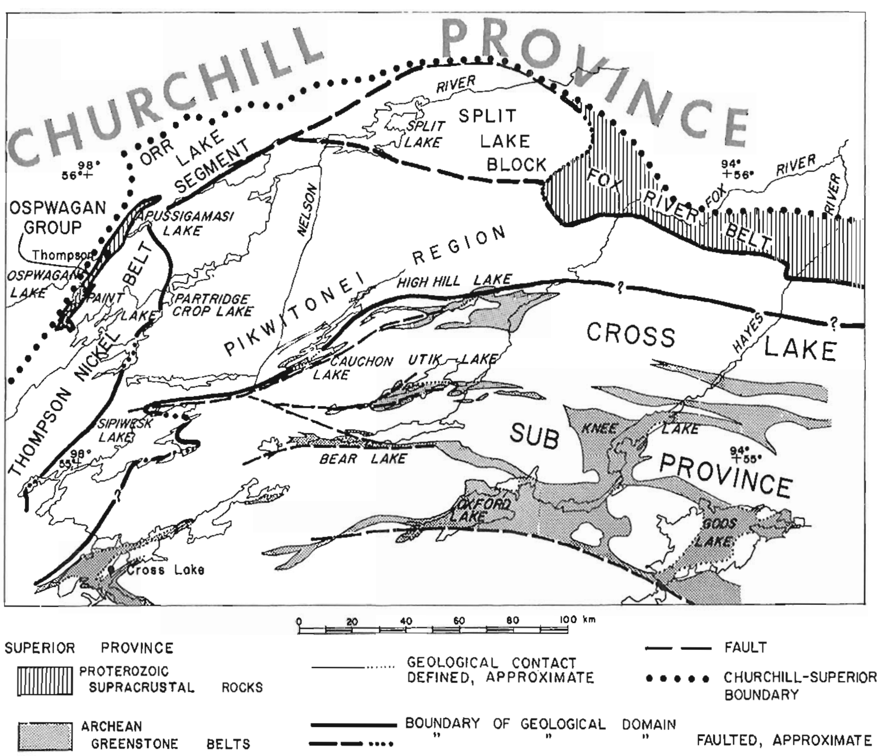

Figure 1. Geological domains of the northwestern Superior Province, Manitoba.

Superior Greenstone Belts

The greenstone belts in the Cross Lake subprovince trend approximately easterly. They are bounded by massive to foliated granitoid intrusive masses and layered tonalitic to granodioritic gneiss, which were derived from granitoid and minor supracrustal rocks. Faults commonly occur along the margins of the belts. In some places faulted greenstone belt margins coincide with steep metamorphic gradients suggesting that these greenstone belts may be downfaulted blocks.[2]

The Oxford Lake-Knee Lake greenstone belt (Fig. 1) contains a typical Archean greenstone belt succession. The Hayes River Group, a lower, predominantly metavolcanic succession, was formed by cyclic basalt-rhyolite volcanism and associated sedimentation (Hubregtse, 1976). The Hayes River Group is unconformably overlain by the Oxford Lake Group, a predominantly metasedimentary succession comprising conglomerate, immature quartzofeldspathic sediments, and minor volcanic rocks of high-K, calc-alkaline and shoshonitic affinity. The two groups are separated by a period of deformation, tonalitic plutonism and metamorphism. The stratigraphy of the Oxford Lake-Knee Lake

[1] Available radiometric ages have not confirmed an early Archean granulite basement (>3000 Ma), as suggested by Bell (1971b), and Ermanovics and Davison (1976). The "basement granodiorite gneiss" at Cross Lake has recently yielded a Kenoran age of 2700 Ma (Ermanovics and Froese, 1978). It is still possible, however, that rocks >3000 Ma make up part of the Pikwitonei region.

[2] These steep gradients occur only over a relatively short distance (2 km or less) and can therefore not be shown on Figure 2.

greenstone belt is essentially repeated in the larger greenstone belts, for example in the Island Lake Belt and Rice Lake Belt south of the area studied.

The greenstone belts north of the Oxford Lake-Knee Lake Belt comprise rocks comparable to the Hayes River Group, but felsic volcanic rocks decrease in abundance and are absent north of Utik Lake. The associated metasedimentary rocks are generally magnesium-rich and reflect a dominant mafic source.

Metamorphism

Volcanogenic metasediment and diagenetically altered, calcium-deficient metabasalt (c.f. Weber, 1974) are among the few lithologic units within the greenstone belts that are compositionally suitable for the formation of mineral assemblages indicative of specific metamorphic conditions. The metamorphic grade generally increases toward the greenstone belt margins. The lowest metamorphic grade registered in successive greenstone belts increases toward the Pikwitonei region granulites (Fig. 2), suggesting that a regional metamorphic gradient is superimposed on local zonations within greenstone belts.

Two metamorphic events have been recognized in the Oxford Lake-Knee Lake greenstone belt, one predates and the other postdates deposition of the Oxford Lake Group (Hubregtse, in prep.). Similarly, two metamorphic events have been identified in the Rice Lake greenstone belt and adjacent Manigotagan gneissic belt in southeastern Manitoba (Weber, 1971; McRitchie and Weber, 1971). In the Oxford Lake-Knee Lake greenstone belt both events range from lower greenschist to lower amphibolite facies. The greenschist facies conditions are characterized by the assemblages chlorite ± carbonate ± quartz ± albite ± muscovite (Hubregtse, in prep.). In the lower amphibolite facies rocks, kyanite was formed during the older event and sillimanite during the younger event, suggesting higher pressure conditions during the pre-Oxford Group metamorphism.

In the Utik and Bear Lake greenstone belts, mineral assemblages do not indicate more than one prograde metamorphism. Assemblages containing andalusite, staurolite, biotite, muscovite, and garnet suggest lower amphibolite facies conditions. Metasedimentary rocks in the central part of the belt contain cordierite, anthophyllite, and biotite with or without garnet. The absence of staurolite in the central part of the belt is probably related to unsuitable rock

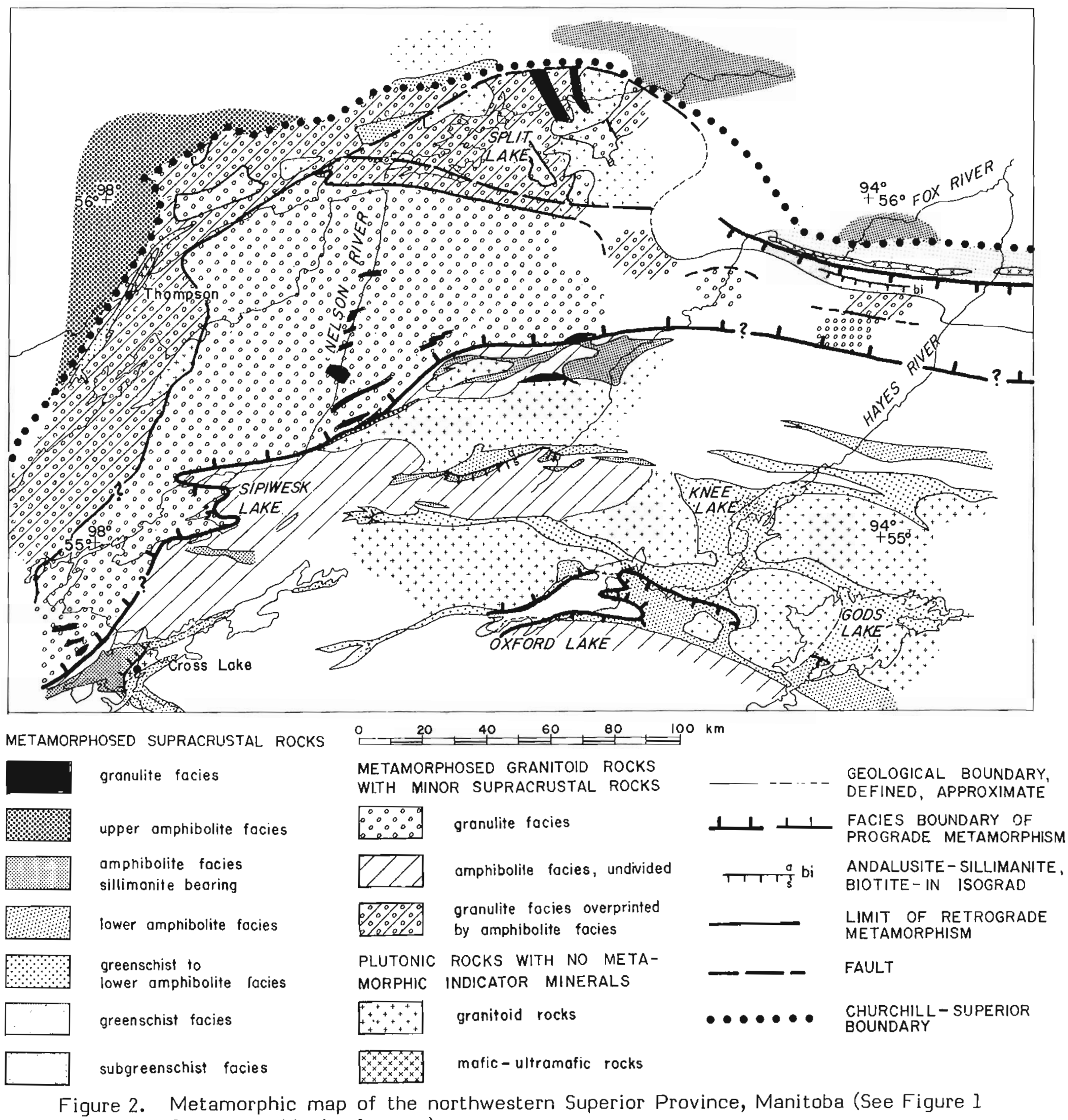

Figure 2. Metamorphic map of the northwestern Superior Province, Manitoba (See Figure 1 for geographical reference).

chemistry. Despite the amphibolite facies metamorphism, primary depositional structures and crystallization textures are remarkably well preserved (Weber, 1974).

An andalusite-sillimanite isograd parallels the southern margin of the Utik Lake greenstone belt (Fig. 2[1]). North of the isograd small sillimanite needles may be associated with staurolite and andalusite. Immediately south of the isograd, sillimanite, partially retrograded to muscovite, occurs in place of staurolite and andalusite. The retrogression is thought to be related to a nearby fault zone whose traces are subparallel to the belt margin. Vertical displacement along this fault may have caused the steep metamorphic gradient along the southern margin of the belt.

South of the andalusite-sillimanite isograd the metavolcanic and associated metasedimentary rocks comprise biotite-bearing amphibolite, biotite-hornblende-plagioclase-quartz-gneiss, hornblende-cummingtonite ± garnet schist, and biotite-sillimanite gneiss. These rocks are increasingly intruded by lits and sills of granodiorite resulting in a 'granitization' of the greenstone belt southwards and its transition into a migmatite terrane. In Figure 2 the migmatite area is shown in "amphibolite facies, undivided", because the boundary between lower and upper amphibolite facies rocks cannot be defined due to the scarcity of muscovite-bearing rocks.

The northern margin of the greenstone belt is intruded by a porphyroblastic granodiorite batholith. No increase of metamorphic grade was noted across the poorly exposed northern greenstone belt margin.

The metamorphic grade in the Utik Lake greenstone belt increases stratigraphically upward (Weber, 1974). Strata in this belt, therefore, were probably overturned prior to regional metamorphism. Thus, this metamorphism is probably equivalent to the post-Oxford Group event.

The Bear Lake greenstone belt comprises metagreywacke and metasiltstone, felsic pyroclastic rocks, and mafic metavolcanic rocks. The absence of alumino-silicate minerals in the metasedimentary rocks appears to be due to their chemical composition. Biotite, muscovite, quartz, and plagioclase ± epidote are the principal minerals. Garnet occurs locally in metabasalt. The assemblages represent metamorphic conditions ranging from upper greenschist to lower amphibolite facies. The contact between the greenstone belt and gneiss to the north is defined by a fault. Migmatitic supracrustal rocks just north of the fault contain sillimanite[2] and cordierite, suggesting that displacement along the fault caused the juxtaposition of high and low grade metamorphic rocks.

Altered mafic rocks in the central part of the High Hill Lake greenstone belt contain cordierite, sillimanite, and biotite suggesting conditions similar to those characteristic of the southern margin of the Utik Lake greenstone belt. Along the northern margin of the High Hill Lake Belt, metabasalt has been recrystallized to orthopyroxene-bearing amphibolite indicating a gradation into hornblende granulite facies. Orthopyroxene is also present in calcic metasedimentary rocks at the southern margin.

The western extension of the High Hill Lake greenstone belt is exposed at Cauchon Lake (Fig. 1). The supracrustal rocks consist of layered amphibolites, mafic, partly calcareous metasedimentary rocks, and quartzite, oxide-silicate iron formation. The layering in the amphibolites is the result of metamorphic segregation of plagioclase from amphibole, superimposed upon a transposed compositional layering which resulted from intensive shearing of pillowed metabasalt. Coarse grained, leuco-gabbro mobilizates form subconcordant lenses and discontinuous layers, parallel to west-northwest trending fold axes.

The southern portion of the belt of supracrustal rocks at Cauchon Lake is in the amphibolite facies. Mafic minerals in the mafic rocks are clinopyroxene and hornblende. The anorthite content of the plagioclase in the amphibolite and mobilizates is 80-85 per cent. We assume that the high anorthite content is the result of high crystallization temperatures (≥700°C, Wenk and Wenk, 1977)[3]. Thus, these supracrustal rocks are shown in the upper amphibolite facies (Fig. 2). In the northern part of the belt, the supracrustal rocks are in granulite facies. The amphibolite contains orthopyroxene, clinopyroxene, and hornblende, and the mobilizate is enderbitic. This suggests that, as at High Hill Lake, the contact between the amphibolite facies and the granulite facies is a prograde transition and is not related to a structural break or a stratigraphic unconformity. The boundary of the Pikwitonei region, i.e. the "orthopyroxene-in" isograd appears to be discordant with the strike of lithologic units, as suggested by recent mapping.

The Pikwitonei Region

The Pikwitonei region is underlain predominantly by granulite facies gneiss and minor amounts of amphibolite facies rocks. Bell (1971b) regarded this gneiss terrane as a basement to the Superior greenstone belts. The main support for this interpretation is an unconformity east of the granulite terrane near the town of Cross Lake (Fig. 1) (120 km south of Thompson) where metasedimentary and metavolcanic rocks of the Cross Lake greenstone belt unconformably overlie granodiorite. Rousell (1965), who discovered the unconformity, also regarded the granodiorite as basement to the greenstone[4]. However, because Rousell was able to map a gradation from amphibolite facies rocks of the Cross Lake greenstone belt into granulite facies rocks to the west (Fig. 2), he concluded that the granulite facies metamorphism postdated the formation of the greenstone belt. Our observations at Cauchon Lake and High Hill Lake confirm Rousell's conclusion.

Approximately 80 per cent of the Pikwitonei region is underlain by granulite facies metaquartz diorite and metagranodiorite, and related migmatitic gneiss. In the Cauchon Lake area the remaining 20 per cent is underlain by discontinuous layers of granulite facies silicate-oxide iron formation, paragneiss, and mafic and ultramafic rocks. Some of the latter are considered to represent supracrustal rocks because they are lithologically similar to rocks of the Utik Lake and High Hill Lake greenstone belts. Supracrustal rocks appear to be lacking in the Sipiwesk Lake area (Hubregtse, 1977) where mafic/ultramafic rocks of deeper-seated origin occur.

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[1] In Figure 2 the isograd coincides with the southern greenstone belt margin, however, in the field the isograd is 1-2 km north of it.

[2] A mineral reported as kyanite (Weber, 1974) has been subsequently identified as sillimanite.

[3] Studies in the Central Alps (Wenk, 1962; Wenk and Keller, 1969; Wenk and Wenk, 1977) have demonstrated that the anorthite content of plagioclase in isochemical rocks increases with increasing temperatures during metamorphism and that in particular the anorthite contents of plagioclase in amphibolite may be suitable for the definition of metamorphic isograds.

[4] Rousell's "basement granodiorite gneiss" is similar to plutons which intrude greenstone belts elsewhere and which are also unconformably overlain by supracrustal rocks in several places, i.e. at Rice Lake (Weber, 1971) and Island Lake (Herd and Ermanovics, 1976). The unconformity at Cross Lake may represent an unconformity within the supracrustal succession and not the base of the greenstone belt. This is supported by a recently obtained U-Pb age of 2700 Ma (Ermanovics and Froese, 1978) for the "basement granodiorite gneiss".

Metamorphism

The earliest metamorphism (M_1) recognizable in the Cauchon Lake area of the Pikwitonei region is associated with a metamorphic layering (S_1) in supracrustal rocks and with lit par lit injections in migmatites. These migmatites were probably formed by partial anatexis of supracrustal rocks and intrusions of tonalite and granodiorite. The garnet and cordierite that formed during the M_1 event in the central part of the Pikwitonei region suggest amphibolite facies conditions (Weber, 1977). In the Sipiwesk Lake area the M_1 event may have reached granulite facies conditions (Hubregtse, 1977).

A period of deformation (D_2) folded the S_1 layering along west-southwest trending axial traces. New S_2 planar structures were developed parallel to those traces during D_2. It is the metamorphism (M_2) during D_2 which reached the granulite facies typical of the Pikwitonei region (Weber, 1977; Hubregtse, 1977).

A third event (M_3), restricted to relatively narrow west-southwest trending zones, resulted in retrogression of the granulite to amphibolite facies assemblages, the formation of augen gneiss, cataclasis and local mylonitization.

Locally porphyroblastic hypersthene[1] with or without syngenetic (?) biotite is a metamorphic indicator of the M_2 metamorphism in leucocratic gneiss and foliated metaplutonic rocks north of Cauchon Lake. The hypersthene may be replaced by clinopyroxene or amphibole through (retrogressive?) reaction with plagioclase. The margin of hypersthene is altered locally to fine grained biotite.

Rocks of intermediate to mafic composition contain hypersthene and/or clinopyroxene, plagioclase with or without garnet, hornblende, biotite, quartz, and magnetite. Ultramafic rocks consist of orthopyroxene with or without olivine, plagioclase, clinopyroxene, hornblende, garnet, spinel, and magnetite. Iron formation and associated metasedimentary rocks contain quartz, garnet, orthopyroxene, magnetite with or without plagioclase and biotite.

Hornblende-granulite facies metamorphism appears to have been prevalent in most of the Pikwitonei region, based on the widespread occurrence of biotite and hornblende; however, on-going mineralogical studies promise to reveal more detailed information regarding the metamorphic history of this area.

Thompson Nickel Belt

The Thompson Nickel Belt is underlain predominantly by migmatitic and layered, cataclastic, felsic gneiss with discontinuous layers and lenses of mafic and minor ultramafic rocks, subconcordant pegmatites and aplites. This suite of rocks named Moak Lake gneiss (Scoates et al., 1977) is interpreted to represent a suite of granulites similar to those of the Pikwitonei region but which have been overprinted by a younger metamorphic-deformational event (Weber, 1976, 1977; Hubregtse, 1977). The Moak Lake gneiss may be garnet-bearing and may contain calc-silicate rocks, and iron formation, rocks which are similar to the supracrustal rocks of the Pikwitonei region. The mineral assemblages indicate a range in metamorphic grade from amphibolite to granulite facies which is largely the result of selective retrogression during the younger overprint of an original granulite facies terrane. Partial anatexis during the younger overprint led to the formation of a wide variety of subconcordant or axial planar granitoid mobilizates, including aplite and pegmatite, which amount to less than 10 per cent of the rocks exposed. The source of water required for partial anatexis of the granulites is not known.

Lower grade metasedimentary, metavolcanic, and ultramafic rocks, the Ospwagan group (Scoates et al., 1977), occur near the western margin of the Thompson Nickel Belt. These supracrustal rocks are intruded by granodiorite. Nickel deposits are associated with ultramafic and metasedimentary rocks of the Ospwagan group and with metasedimentary rocks within Moak Lake gneiss whose association with Ospwagan group rocks is not yet known.

The contact between Moak Lake gneiss and Aphebian metagreywacke and metashale strata of the Kisseynew Sedimentary Gneiss Belt marks the western limit of the Thompson Nickel Belt.

Moak Lake gneiss grades into Pikwitonei granulite along the eastern margin of the Thompson Nickel Belt. This transition zone marks the eastern limit of a structural and metamorphic overprint affecting Pikwitonei granulite. In the transition zone at Partridge Crop Lake and Apussigamasi Lake the rocks have been overprinted by tight folds generated about north-northeast-trending axial planes. This has resulted in transition of the metamorphic layering (S_2 – recognized in the Pikwitonei region) into north-northeast-striking planar structures and recrystallization of the granulite facies rocks under amphibolite facies conditions. This transitional contact was also described by Cranstone and Turek (1976, p. 1063) and is indicated by Patterson (1963). The transition is partly obliterated by the intrusion of postgranulite granodiorite (Harrison, 1950; Weber and Scoates, 1976; Hubregtse et al., 1977) and associated metasomatism which has led to the formation of microcline augen gneiss near the intrusions.

In the main part of the Thompson Nickel Belt, small migmatitic zones display the lithologic and structural features and granulite facies metamorphism similar to the S_2 domains of the Pikwitonei region. These zones apparently were partially preserved from the structural and metamorphic overprint that affected the Thompson Nickel Belt.

Metamorphism

Mineral reactions indicative of the metamorphic conditions during the retrogressive overprint in the Apussigamasi and Partridge Crop lakes area include:

$$\text{clinopyroxene} \rightarrow \text{hornblende} \qquad (R.1)$$

$$\text{garnet} \rightarrow \text{hornblende and} \qquad (R.2)$$

$$\text{orthopyroxene} + \text{plagioclase} \rightarrow \text{hornblende} + \text{quartz} \qquad (R.3)$$

Reaction (3) commonly produced a myrmekitic quartz-hornblende intergrowth. Metasomatism related to the younger granitic intrusions and pegmatites led to the formation of biotite replacing clinopyroxene, garnet, or hornblende, the alteration of garnet to biotite with or without sillimanite, and the formation of cummingtonite from orthopyroxene.

Coats et al. (1972) and Hubregtse (1977) interpreted certain granulite at Paint Lake and a few kilometres to the southwest, near Wintering Lake as indicating that the younger overprint may have achieved granulite facies conditions. Alternatively, these rocks represent unretrogressed Pikwitonei granulite. Current mapping will deal with this problem.

Split Lake Block

The term "Split Lake Block" (Fig. 1) is herein used to refer to the area bounded by the Split Lake fault zone and Assean Lake fault zone in the south and northwest,

[1] The composition of some orthopyroxenes was determined optically.

respectively (Haugh, 1969; Hubregtse, 1975) and by the Gull Rapids fault zone on the Nelson River to the north (Corkery, 1975). The eastern boundary is drawn on the basis of aeromagnetic signatures.

The Split Lake Block (Figs. 1, 2) is underlain by felsic and mafic granulite, retrogressed granulite, and amphibolite facies rocks (Hubregtse, 1975; Haugh, 1969; Corkery, 1975, 1977). The granulites are similar to those in the Pikwitonei region and the amphibolite facies rocks are comparable to the Moak Lake gneiss of the Thompson Nickel Belt. Younger granitoid masses occur near the margins of the Split Lake Block.

The Orr Lake Segment (Fig. 1) appears to share many of the characteristics of the Split Lake Block (McRitchie, 1977b), but is not as well documented because of poor exposure.

AGE OF METAMORPHISM IN THE SUPERIOR PROVINCE

Biotite K-Ar ages in the greenstone belt terrane of the Cross Lake subprovince are unlike those in most of the Superior Province and range from 1600 to approximately 2600 Ma. Biotite and muscovite ages from metasedimentary rocks of the Cross Lake and Utik Lake greenstone belts range from 1680 to 2160 Ma (6 ages) suggesting crystallization of the micas at 2160 Ma or earlier, most likely during the Kenoran orogeny, more than 2500 Ma ago, and subsequent argon loss during one or more events probably not younger than 1600 Ma. Stockwell et al. (1970) assumed that the argon loss was caused by heat effects of the Hudsonian Orogeny.

Available Rb-Sr and U-Pb age determinations indicate a late-Archean age for the greenstone belts in the Cross Lake subprovince. A preliminary Rb-Sr whole rock isochron age of 2460 Ma[1] (Cheung, 1978) was obtained from a granodiorite which is intrusive into Oxford Lake Group rocks.[2] This is interpreted as a minimum age for the intrusion and is, therefore, a minimum age for the deposition of the Oxford Lake Group.

Zircons from Hayes River Group metadacite give a U-Pb age of 2700 Ma (E.J. Catanzaro, pers. comm., 1977). Since the U-Pb data are relatively discordant, 2700 Ma is considered a minimum age for the metadacite (G.N. Hanson, pers. comm.), and it is thus a minimum age for the deposition of the Hayes River Group. U-Pb data from Superior Province greenstone belts farther south (Noranda, Uchi Lake) yield slightly older ages (2755-2800 Ma) for synchronous volcanism and plutonism (Krogh and Davis, 1971).

The 2460 Ma Rb-Sr age for the granodiorite which is intrusive into rocks of the Oxford Lake Group and two additional preliminary Rb-Sr whole rock ages from the Gods Lake-Goose Lake area indicate a regional metamorphic-plutonic event at between 2400 and 2500 Ma (Cheung, 1978). This event is considered to be equivalent to the younger metamorphic event recognized in the Oxford Lake area which in turn is similar in age to a corresponding event in the Rice Lake area in southeastern Manitoba (Uchi Belt), dated at 2480 Ma (Turek and Peterman, 1971). This metamorphic-plutonic event appears to mark the end of the Kenoran Orogeny in the western Superior Province.

U-Pb ages from the Berens Belt, south of the Cross Lake subprovince, fall into two groups, 2690-2760 Ma, and 2900-3000 Ma indicating two distinctly different orogenic episodes. These might correlate with the 2650 Ma and 3000 Ma orogenic episodes of the English River Belt (Krogh et al., 1976). In the Rice Lake greenstone belt Rb-Sr ages yield 2680 Ma for the oldest recognizable metamorphic-plutonic event (Turek and Peterman, 1971) which appears to correspond to the younger of the two orogenic episodes in the adjacent granitoid terranes. Available data thus suggest, that a widespread Kenoran plutonic and metamorphic event took place between 2650 and 2760 Ma in the western Superior Province. The older metamorphic event recognized in the Oxford Lake area is considered to be part of this Kenoran event.[3]

Available radiometric ages for granulite of the Pikwitonei region suggest an Archean age for the granulite facies metamorphism. Published K-Ar biotite ages range from 2400 to 2700 Ma. The range in age may reflect argon loss subsequent to the formation of biotite, may be due to more than one biotite population, or may be due to a combination of these factors. Most biotite appears to coexist with orthopyroxene and such biotite would yield a K-Ar age which would be a minimum age for the granulite facies metamorphism. However, in some rocks hypersthene is replaced by younger biotite, resulting in two biotite populations. Biotites from those rocks would give younger K-Ar ages.

Assuming argon loss in similar proportions as in the Cross Lake subprovince greenstone belts, Bell (1971b) postulated an age greater than 3000 Ma for the granulite of the Pikwitonei region. Ermanovics and Davison (1976) also assumed an age older than 3000 Ma for the Pikwitonei granulite, essentially using Bell's arguments, despite the lack of any supporting Rb-Sr or U-Pb ages from this area. Rb-Sr age studies in this area are now in progress.

If granulite facies supracrustal rocks in the Pikwitonei region are of the same age as the greenstones of the Oxford Lake-Knee Lake area, then the granulite facies metamorphism must be younger than those greenstones, i.e. <2800 Ma. This does not preclude the possibility that rocks older than 2800 Ma occur in the Pikwitonei region, as in the similar Berens Belt and the English River Belt. However, whether an apparent lithologic similarity among these belts is the result of similar age or the result of a similar succession of geological events remains to be tested.

Published biotite ages from gneiss of the Thompson Nickel Belt show a narrow range of 1700 to 1800 Ma. These ages are generally regarded as reflecting a Hudsonian event which affected this area (Stockwell et al., 1970). Rance (1966) and Bell (1966, 1971b) regarded the Hudsonian Orogeny as the principal overprinting event on the basis of field data. However, Cranstone and Turek (1976) considered the major overprint to be of Archean age. They interpreted a Rb-Sr whole rock age of 2760 Ma from "Wabowden migmatitic gneiss" as dating "formation of the gneisses by remobilization of the Pikwitonei granulites". An age of 1680 Ma obtained from two samples of the same gneiss was interpreted as the result of "some Hudsonian overprinting". Cranstone and Turek concluded that "the Kenoran orogeny was the dominant tectonothermal event and that the pronounced northeasterly structural trend developed during the Kenoran orogeny. Additional northeast trending tectonic deformation (refolding and faulting) and metamorphism occurred during the Hudsonian orogeny".

The northeast trends of fold axes and metamorphic layering associated with the younger metamorphic overprint are apparently related to the parallel trends in the adjacent metasedimentary gneiss of the Churchill Province west of Setting Lake (e.g. Rance, 1966).[1] Rb-Sr data (Clark et al.,

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[1] All Rb-Sr isochron ages have been (re-)calculated with the ^{87}Rb decay constant of $1.42 \times 10^{-11} yr^{-1}$.

[2] Biotite from this granodiorite has yielded a K-Ar age of 1650 Ma (Moore et al., 1960) supporting the assumption of Ar loss.

[3] The 2700 Ma granodiorite at Cross Lake (Ermanovics and Froese, 1978) is considered to have intruded during this event and 2700 Ma is, therefore, an approximate age for this event in the Cross Lake subprovince.

1974) indicate that the metasedimentary gneiss was derived from Aphebian igneous rocks and that this gneiss had undergone high grade metamorphism during the Hudsonian Orogeny at 1700-1800 Ma. It appears reasonable on the basis of the foregoing data to postulate a 1700-1800 Ma age for the metamorphic and structural overprint in the Thompson Nickel Belt. The K-Ar biotite ages from gneiss of the Thompson Nickel Belt indicate this event. The 1680 Ma and 1760 Ma Rb-Sr ages by Cranstone and Turek (1976) for the "Wabowden migmatitic gneiss" and for a pegmatite from Moak Lake probably reflect the same period of recrystallization.

We consider Cranstone and Turek's age of 2780 Ma for the "Wabowden migmatitic gneiss" as dating the preceding granulite facies metamorphism. It is more likely (e.g. Doig, 1977) that the Rb-Sr whole rock isochron age of an overprinted granulite dates the older metamorphism (or an original plutonic rock) rather than an overprint of lower metamorphic grade. Specific sampling may identify the age of a younger event, if localized isotope homogenization has taken place during that event (e.g. Hänny et al., 1975). Such an event is apparently represented by Cranstone and Turek's data yielding the 1680 Ma age for the "Wabowden migmatitic gneiss".

FOX RIVER BELT AND OSPWAGAN GROUP

Rocks of the Fox River Belt (Scoates, 1975a, 1977, in prep.) and Ospwagan group (Scoates et al., 1977) are lithologically similar. In turn, these rocks are similar to rocks of the Circum-Ungava geosyncline, in particular, those in the Cape Smith-Wakeham Bay belt (Baragar, 1975; Fujiwara and Schwarz, 1975; Stam, 1961; and Taylor et al., 1975). The Fox River Belt occupies the junction between the Superior and Churchill provinces, and the Ospwagan group forms folded structures in Moak Lake gneiss near the western margin of the Thompson Nickel Belt.

Although poorly exposed, the Fox River Belt is presently better known than the Ospwagan group. It is less deformed and has undergone a lower grade of metamorphism and this allows for a more straightforward interpretation of data. Most information has been obtained from drill core from the western 50 km of the Fox River Belt. The core was generously provided by the International Nickel Company of Canada Limited for detailed studies.

The Fox River Belt comprises a sequence of sedimentary rocks, mafic to ultramafic komatiitic and tholeiitic volcanic rocks and mafic and ultramafic intrusive rocks that form a north-facing homoclinal belt, which is interpreted to be resting on Superior Province gneiss. Table 1 shows a stratigraphic cross section with metamorphic grades and indicative mineral assemblages.

The Ospwagan group of the Thompson Nickel Belt coincides largely with units 5, 6 and 11 of Quirke et al. (1970). Rocks of the Ospwagan group form a continuous belt from Ospwagan Lake to Moak Lake north of Apussigamasi Lake (Fig. 1). The stratigraphy, structure, and extent of the Ospwagan group is a major topic of the Nickel Belt Project initiated in 1976 by the Manitoba Geological Survey Section.

Metamorphism

Prehnite and pumpellyite, indicative of subgreenschist facies metamorphic conditions, occur in the komatiitic and tholeiitic volcanic rocks in the stratigraphically upper part of the Fox River Belt (Fig. 2). Prehnite occurs as irregular patches replacing plagioclase and is associated with quartz, chlorite and epidote, and carbonate veinlets. Pumpellyite replaces plagioclase and occurs as irregular, very fine grained patches associated with chlorite and epidote. Sphene pseudomorphously replaces ilmenite. The stratigraphically lower half of the Fox River Belt contains chlorite, tremolite, epidote and sphene, the same metamorphic minerals that are present in the upper volcanic units. However, pumpellyite and prehnite are absent in the lower section, suggesting greenschist facies metamorphic conditions. Toward the base of the Lower Volcanic formation, primary clinopyroxene is increasingly altered to tremolite. Biotite and locally chloritoid are developed in the Lower Sedimentary formation. The metamorphic mineral assemblages observed within Fox River Belt stratigraphy are consistent with temperature/pressure conditions in terms of estimated burial. The increase in metamorphic grade toward the base of the Fox River Belt is thus interpreted as being due to burial.

It is not yet possible to describe in detail the metamorphic history of Ospwagan group rocks. Preliminary work indicates a range in metamorphic grade from greenschist to amphibolite facies. Contact metamorphic effects are superimposed on the regional metamorphic variations in Ospwagan group rocks adjacent to felsic plutons. Continuing examination of Ospwagan group rocks will significantly enhance our knowledge concerning their metamorphism.

Age of Fox River Belt and Ospwagan Group

The rocks of the Fox River Belt contrast distinctly with those of the adjacent Churchill and Superior provinces, with regard to lithology, intensity of deformation, and grade of metamorphism. The Fox River Belt has undergone burial metamorphism and, with the exception of tilting into a vertical position, is not deformed. This belt is interpreted as overlying Superior Province gneiss which is, at least partially, retrogressed granulite (Scoates et al., 1977) and is flanked on the north side by lower to upper amphibolite facies Kisseynew-type gneiss. The Ospwagan group, although more deformed and metamorphosed, is lithologically similar to the Fox River Belt.

The geological evidence suggests that the rocks of the Fox River Belt are younger than the adjacent rocks. If the age of metasedimentary gneiss of the Churchill Province is Aphebian and if these rocks were metamorphosed during the Hudsonian Orogeny (1800 Ma), as is suggested by radiometric ages, the Fox River Belt could be younger. In support of this supposition is a preliminary Rb-Sr whole rock isochron age of 1610 Ma (Scoates and Clark, in prep.) which has been obtained from hornfelsed quartz-rich siltstone below and immediately adjacent to the Fox River Sill. This age probably dates the intrusion of the sill, as the hornfels is the result of contact metamorphism. A direct relationship between intrusions and lava of the Fox River Belt is suggested by a continuum of chemical parameters (Scoates, in prep.). The volcanic rocks overlying the Fox River Sill, therefore, are considered to be equivalent in age to the sill.

Ospwagan group rocks have been deformed and metamorphosed subsequent to their deposition. As previously noted these rocks are similar to Fox River Belt rocks. A number of possibilities exist in establishing age relations, and three are suggested:

1) Ospwagan group rocks are older than Fox River Belt rocks and have been subjected to the Hudsonian Orogeny;
2) Ospwagan group and Fox River Belt rocks are the same age and predate the Hudsonian Orogeny and Fox River Belt rocks have somehow escaped the effects of that event; and
3) Ospwagan group and Fox River Belt rocks are the same age and postdate the Hudsonian Orogeny – and only the Ospwagan group has been affected by a post-Hudsonian metamorphic and deformational event.

[1] Recent structural analyses indicate just one phase of pronounced northeasterly folding (Hubregtse, 1977) and not two, as proposed by Cranstone and Turek (1976).

Table 1

Stratigraphy and Metamorphism of Fox River Belt

Formation		Thickness (km)	Lithology	Metamorphic Grade	Typical Mineral Assemblages	Preliminary Rb-Sr whole rock age (Ma)[1]
Aphebian gneiss of the Churchill Province			Garnetiferous biotite and hornblende gneiss	Lower Amphibolite	garnet, biotite, staurolite, sillimanite, kyanite, muscovite	1740
			— Contact not exposed fault(?) —			
Fox River Belt	Upper Sedimentary formation	1.0-2.0	Argillite, shale			
			— Contact not exposed —			
	Upper Volcanic formation	3.0-3.5	Pillowed and massive pyroxenitic and basaltic komatiite, some layered flows, massive and pillowed tholeiitic basalt	Subgreenschist	chlorite, tremolite, epidote, sphene, ± prehnite, pumpellyite	
			— Contact not exposed —	Increasing metamorphic grade ↓		
	Middle Sedimentary formation	0.1-0.4	Quartz-bearing siltstone			
	Fox River Sill	2.0				
			— Intrusive Contact —			
	Middle Sedimentary formation	0.2-0.4	Quartz-bearing siltstone			1610
			— Contact not exposed —			
	Lower Volcanic formation	2.0-2.5	Pillowed and massive pyroxenitic and basaltic komatiite, some layered flows, massive and pillowed tholeiitic basalt	Greenschist	chlorite, tremolite, epidote, sphene	
			— Contact not exposed —			
	Lower Sedimentary formation	4.0-4.5?	Siltstone, argillite, marl, quartzite, and iron formation, some differentiated intrusions in upper part		biotite, chloritoid	
			- Contact not exposed fault(?) —			
Archean gneiss of the Superior Province			Gneissic granite and layered gneiss	Amphibolite (local retrogressed granulite)	biotite, hornblende	2679

[1] Scoates, R.F.J. and Clark, G.S. (in prep.).

The latter possibility is supported by a Rb-Sr whole rock isochron age of 1615 Ma (Cranstone and Turek, 1976) obtained from Ospwagan group quartz-rich argillite schist. We concur with Cranstone and Turek's first alternative explanation for this age that it reflects a metamorphic event possibly related to the intrusion of the Mystery Lake granodiorite, as supported by 1600 Ma biotite and muscovite K-Ar ages (Wanless et al., 1968) obtained from that intrusion. Further support for this possibility is found in four biotite and muscovite K-Ar ages determined from the Thompson mine which range from 1515 ± 50 Ma to 1630 ± 50 Ma (Wanless et al., 1968).

CHURCHILL PROVINCE

Greywacke- and shale-derived gneiss, comparable to gneiss of the Kisseynew Sedimentary Gneiss Belt (Bailes and McRitchie, 1978), underlies the Churchill Province adjacent to the Churchill-Superior boundary (Fig. 2). Garnet-cordierite-sillimanite-biotite assemblages are reported from this gneiss west of Ospwagan Lake (Stephenson, 1974) suggesting upper amphibolite facies conditions. Further west the metamorphic grade increases to hornblende granulite facies (Bailes and McRitchie, 1978). McRitchie (1977b) described garnet-cordierite-sillimanite-K-feldspar assemblages, also indicating upper amphibolite facies conditions, from the Harding-Rock Lakes area west of the Orr Lake Segment.

Assemblages containing garnet ± sillimanite ± staurolite ± kyanite + muscovite occur in gneiss north of the Fox River Belt (McRitchie, 1977a), suggesting lower to middle amphibolite facies conditions (Table 1). Haugh and Elphick (1968)[1] and Frohlinger (1974) reported garnet ± sillimanite ± cordierite from gneiss northeast of the Gull Rapids fault zone indicating metamorphic conditions similar to those of the Churchill Province gneiss along the Fox River.

Rb-Sr whole rock data (Clark et al., 1974; Scoates and Clark, in prep.) strongly suggest that this metasedimentary gneiss was formed during the Hudsonian Orogeny approximately 1800 Ma ago. Low $^{87}Sr/^{86}Sr$ initial ratios suggest that the original sediments were derived from Aphebian and not Archean igneous rocks.

THE CHURCHILL-SUPERIOR BOUNDARY

The contact between Aphebian greywacke- and shale-derived Kisseynew-type gneiss and Archean, Pikwitonei-type gneiss is considered to be the most fundamental contact of the boundary zone in Manitoba and is herein defined as the Churchill-Superior boundary. Although this contact has been recognized as separating Aphebian and Archean rocks in certain areas (Rance, 1966; Bell, 1971b), its fundamental nature has only been appreciated during recent mapping along the boundary zone. Structural discordance between Archean and Aphebian gneiss occurs at this boundary along the western edge of the Thompson Nickel Belt (Rance, 1966; Bell, 1971b; McRitchie, 1977b), the northwestern margin of the Orr Lake Segment (McRitchie, 1977b), and the northern margin of the Split Lake Block (Gill, 1948; Corkery, 1977). Aphebian and Archean gneisses are lithologically and chronologically distinct.

The Gull Rapids fault zone is the boundary along the north margin of the Split Lake Block (Fig. 1) and the boundary is a fault between Thompson and Ospwagan Lake. The nature of the boundary along the western edge of the Orr Lake Segment is not known. The northeastern extension of the boundary is buried beneath Fox River Belt supracrustal rocks. It is not known whether Aphebian gneiss is in contact with Archean gneiss at depth; however, if such were the case the boundary would project somewhere within the Fox River Belt. In view of this uncertainty the boundary is positioned along the northern contact of the Fox River Belt.

The original nature of this Aphebian-Archean contact is not known because of substantial deformation associated with the boundary zone. However, it is reasonable to infer that the Aphebian gneiss belonged to a supracrustal sequence which was, in part, deposited unconformably upon the edge of the Archean craton. Stockwell et al. (1970) stated, "Boundaries between (structural) provinces are drawn where one trend is truncated by another, either along major unconformities or, in their absence along orogenic fronts." From the preceding discussion it is clear that the Aphebian-Archean contact is the most fundamental contact within the Churchill-Superior boundary zone. It is the only contact which satisfies Stockwell's definition of structural province boundaries.

Bell (1971b) suggested that the contact between the Thompson Nickel Belt and the Pikwitonei region be considered the edge of a Hudsonian Front which had overprinted the margin of an Archean craton.[2] This contact has been defined as the Churchill boundary by Bell (1971b) and on the Tectonic Map of Canada (Geological Survey of Canada, 1969). However, this contact could only be defined as the boundary if the fundamental Aphebian-Archean contact already described did not exist. Furthermore, a structural province boundary along the eastern margin of the Thompson Nickel Belt is impractical because of the gradational nature of that contact, whereas the Aphebian-Archean contact is more definitive.

SUMMARY AND CONCLUSIONS

An increase in the metamorphic grade is noted near some of the greenstone belt margins in the northwestern corner of the Superior Province (Cross Lake subprovince). This appears to be the result, in part, of down-faulting of the greenstone belts.

Most mineral assemblages in the greenstone belts of the Cross Lake subprovince are considered to have been formed during the second of two Archean metamorphic events. The older event, recognized only in the Oxford Lake area, predated the deposition of the Oxford Lake Group and is characterized by higher pressure conditions than the younger post-Oxford Lake Group event. Radiometric data suggest that both metamorphic events occurred between 2400 and 2800 Ma.

The lowest metamorphic grade registered in successive greenstone belts increases towards the granulite facies rocks of the Pikwitonei region suggesting a regional metamorphic gradient. An apparent decrease, in the same direction, in the volume of units typical for the stratigraphically upper section of Superior Province greenstone belt successions (e.g. Oxford Lake-Knee Lake greenstone belt), suggests that from southeast to northwest and north, successively deeper crustal levels may be exposed.

The contact between the greenstone belt terrane and the Pikwitonei region is a prograde transition from amphibolite facies to granulite facies and is, therefore, not an unconformity as inferred by Bell (1971b). Among the oldest rocks in the Cauchon Lake area of the Pikwitonei region are granulite facies supracrustal rocks which are lithologically similar to parts of greenstone belt successions that are found farther south. If these supracrustal rocks are the equivalent of Superior Province greenstone rocks, then the granulite facies metamorphism is younger than those greenstone rocks, i.e. younger than 2800 Ma and not older as suggested by Bell (1971b).

[1] A mineral reported as kyanite was subsequently identified as sillimanite.

[2] However, Cranstone and Turek (1976) consider the overprint to be of Archean age.

Table 2

Summary of geological events in the western Superior Province and the adjacent portion of the Churchill Province

Tectonic entities	Ages (Ma) of supracrustal rocks and metamorphic (-plutonic) events (dashed lines)							
	1600	1800	2000	2200	2400	2600	2800	3000
Churchill Province (Central Manitoba)		A sv						
Superior Province (Western part)					KENORAN			
Cross Lake subprovince								
– Fox River Belt	S^+ vs							
– Ospwagan group	A^- vs							
– Thompson Nickel Belt								
– Orr Lake Segment		A			? ←	G v		
– Split Lake Block								
– Pikwitonei region		(A)			? ←	GA v		
– Oxford-Knee-Gods lakes (Sachigo Belt)		HUDSONIAN			A^- s	A^- v		
Berens Belt					(A)	A v		A
Rice Lake (Uchi Belt)					G^+ s	G^+ v		
English River Belt					(A)	A^+ s		A

EXPLANATION

Metamorphic facies		Supracrustal rocks	
G	Granulite	s	Sedimentary rocks predominant
A	Amphibolite	v	Volcanic rocks predominant, minor sedimentary rocks
A^+	Amphibolite, minor granulite		
A^-	Amphibolite to greenschist		
G	Greenschist		
G^+	Greenschist, minor amphibolite		
S	Subgreenschist		
S^+	Subgreenschist, minor greenschist		
( )	Minor event		

The Pikwitonei region grades into the Thompson Nickel Belt along the western border of the Superior Province. The Thompson Nickel Belt is underlain predominantly by amphibolite facies gneiss which was formed by a metamorphic and structural overprint of granulites similar to those of the Pikwitonei region. This overprint is considered to be due to the 1800 Ma Hudsonian Orogeny, and may be interpreted as a Hudsonian Front. The Orr Lake Segment and Split Lake Block along the northern margin of the Superior Province although less intensely overprinted, share the geological history of the Thompson Nickel Belt.

Upper greenschist to lower amphibolite facies Ospwagan group supracrustal rocks form fold structures near the western margin of the Thompson Nickel Belt. A similar suite of rocks, the Fox River Belt, occurs along the northern boundary of the Superior Province east of Split Lake. The rocks of the Fox River Belt are essentially undeformed and have undergone subgreenschist to greenschist facies metamorphism. Radiometric ages are interpreted as suggesting a Paleohelikian age of 1610 Ma for the formation of these rocks. The low grade metamorphism is assumed to be of similar age.

The Churchill Province is underlain by metagreywacke and metashale strata of the Kisseynew Sedimentary Gneiss Belt along the Churchill-Superior boundary in Manitoba. Mineral assemblages suggest that these rocks have undergone lower to upper amphibolite metamorphism. Rb-Sr isochron data indicate an age of 1740-1790 Ma for the metamorphism which is also an age for the Hudsonian Orogeny in Manitoba.

Table 2 summarizes the main geological events in the Superior Province and the adjacent Churchill Province of Manitoba.

Conclusions based on new information presented in this paper allow for a better definition of the Churchill-Superior boundary in Manitoba. It approximates the boundary suggested by Stockwell et al. (1970) except between Split Lake and Thompson where it has been repositioned farther north.

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METAMORPHISM OF THE SUPERIOR PROVINCE IN MANITOBA

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Abstract

The Superior province has been divided into several subprovinces, defined on the basis of lithology and structure. The Sachigo, Uchi, and Wabigoon subprovinces consist of volcanic belts intruded by concordant gneissic batholiths and by unmetamorphosed massive plutons. The metamorphic grade in the supracrustal rocks varies from greenschist to lower amphibolite facies, increasing, in some localities, towards the contact with the surrounding intrusions. In addition to such variations, a regional increase in metamorphic grade is present in the Sachigo subprovince apparently leading to a gradational contact with granulites of the Pikwitonei subprovince. The same type of metamorphism observed in the Sachigo, Uchi, and Wabigoon subprovinces appears to be developed also in the Berens subprovince which contains only a few remnants of supracrustal rocks. By way of contrast, the English River subprovince, characterized by metasedimentary gneisses, displays a well-developed metamorphic zonation over a short distance, indicating steep metamorphic gradients. Most of this subprovince is metamorphosed in the upper amphibolite facies giving rise to anatexis in pelitic rocks. This style of metamorphism appears to be related to a somewhat thinner crust. The occurrence of cordierite and andalusite in rocks of the lower amphibolite facies in the Sachigo, Uchi, and Wabigoon subprovinces as well as in the English River subprovince indicates relatively low pressures of metamorphism.

Résumé

On a partagé la province du lac Supérieur en plusieurs sous-provinces, définies d'après la lithologie et la structure. Les sous-provinces de Sachigo, Uchi et Wabigoon sont constituées de zones volcaniques dans lesquelles ont pénétré des batholites gneissiques concordants et des plutons massifs non métamorphisés. Dans les roches supracrustales, l'intensité du métamorphisme varie de l'intensité du faciès schistes vertes à celle du sous-faciès amphibolite inférieure; dans certaines localités, le métamorphisme augmente d'intensité à mesure qu'on se rapproche de contact avec les intrusions environnantes. Outre ces variations, on observe un accroissement régional de l'intensité du métamorphisme dans la sous-province de Sachigo, indiquant apparemment un contact graduel avec les granulites de la sous-province de Pikwitonei. Le type de métamorphisme observé dans les provinces de Sachigo, Uchi et Wabigoon semble aussi se manifester dans la sous-province de Berens, qui ne contient que quelques lambeaux de roches supracrustales. Au contraire, la sous-province de English River, caractérisée par la présence de gneiss métasédimentaires, manifeste une zonalité métamorphique bien développée sur une courte distance, ce qui indique des gradients métamorphiques prononcés. La majeure partie de cette zone est métamorphisée dans le sous-faciès amphibolite supérieure, et ce métamorphisme a provoqué l'anatexie des roches pélitiques. Il semble que ce style de métamorphisme soit dû à l'amincissement de la croûte dans la région. La présence de cordiérite et d'andalousite dans les roches du sous-faciès amphibolite inférieure, dans les provinces de Sachigo, Uchi et Wabigoon, ainsi que dans la sous-province de English River indique que ce métamorphisme s'est effectué à des pressions relativement basses.

INTRODUCTION

The tectonic setting of the subprovinces of the Superior province represented in Manitoba has been discussed by Stockwell (1964), Bell (1971), Ermanovics (1971), Wilson (1971), Ermanovics and Davison (1976), and Beakhouse (1977). This paper describes the nature of metamorphism in each subprovince and examines the relationship of metamorphism to other aspects of geology. Except for using the term subprovince instead of belt, the subdivision of the Superior province proposed by Douglas (1973) is followed:

Douglas (1973):	This paper:
Sachigo volcanic belt	Sachigo subprovince
Pikwitonei belt	Pikwitonei subprovince
Berens plutonic belt	Berens subprovince
Uchi volcanic belt	Uchi subprovince
English River gneiss belt	English River subprovince
Wabigoon volcanic belt	Wabigoon subprovince

The metamorphism of the Pikwitonei subprovince and part of the Sachigo subprovince, discussed by Weber and Scoates (1978), is also considered in this paper in order to provide an account of metamorphism in the Superior province. The nature and location of the Churchill-Superior boundary have been disputed for some time (Bell, 1971; Weber and Scoates, 1978). To be consistent with the structural subdivisions given by Douglas (1973), we have accepted the boundary proposed by Bell (1971). However, we wish to draw attention to a discussion of the boundary problem by Weber and Scoates (1978).

The documentation of sources requires a comment. For the most part, only latest publications have been cited; older studies used are not listed in the references if they appear in later publications referred to in this paper. Much information has been taken from geological reports accompanying maps at a scale of one inch to four miles (Fig. 1). In the southern part, such studies commonly represent the latest geological work. However, in the northern part, the four mile maps are old and sketchy. More recent detailed work is available as a result of

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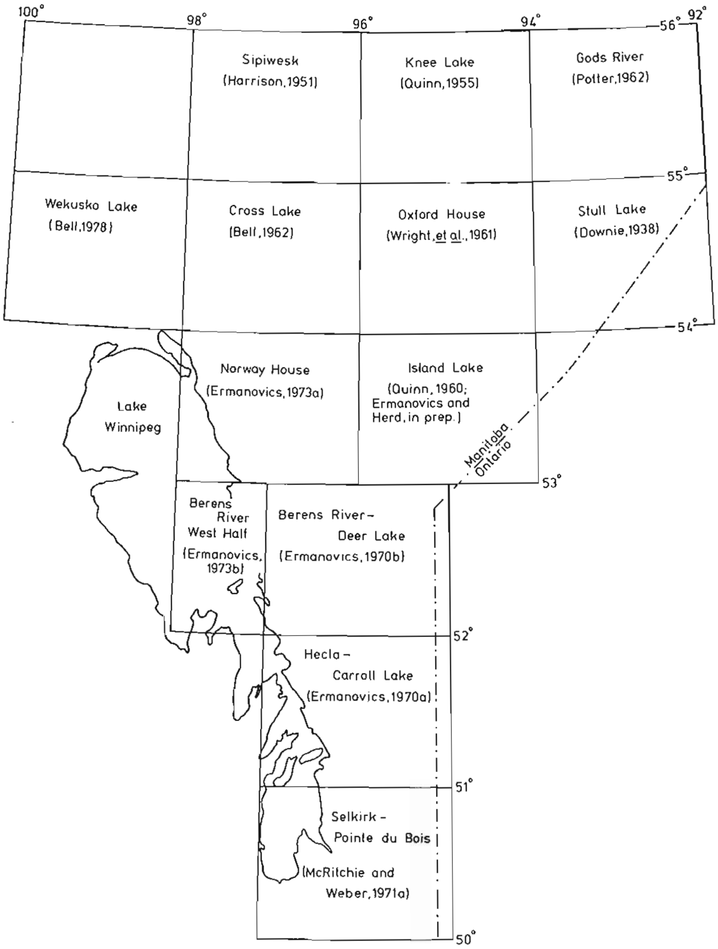


Figure 1. Location of areas mapped on a scale of one inch to 4 miles.

the "Greenstone Project" (Elbers and Gilbert, 1972; Elbers et al., 1973), and subsequent work by geologists of the Manitoba Mineral Resources Division. References to these investigations are given by Weber and Scoates (1978). In the southern part, we have used the reports of "Project Pioneer" (McRitchie and Weber, 1971a) and recent studies in south-eastern Manitoba (Janes and Malyon, 1977).

DEFINITIONS OF METAMORPHIC GRADES

The divisions of metamorphic grade given by Winkler (1976) are used; however, the term facies and the well-established facies names are retained. A correlation of divisions of metamorphic grade used in recent textbooks is given in Table 1. In this paper, a facies is considered to comprise rocks equilibrated within a specified volume of P-T-a_{H_2O} space. The activity of water a_{H_2O} is defined in terms of the chemical potential $\mu - \mu^* = RT \ln a$, where μ^* is the chemical potential of pure water vapour at any given temperature and pressure. The activity has a maximum value of one; it may be lower due to constituents other than water being present in the vapour phase or due to the absence of a vapour phase.

If the number of solid phases involved in a dehydration reaction is equal to the number of components required to describe the phases, the chemical potential of each component, including that of H_2O, is defined at fixed pressure and temperature. Consequently, such reactions may be represented as surfaces in P-T-a_{H_2O} space. In many instances, it is possible to map separate areas of two alternate mineral assemblages related by a dehydration reaction. The corresponding boundary is defined as an isograd based on a specific reaction. A regular pattern of isograds suggests that a_{H_2O} does not change rapidly over short distances (Fyfe and Turner, 1958). Commonly such regularity is taken as an indication that $P_{H_2O} = P_{total}$, i.e. $a_{H_2O} = 1$. However, to account for the regular pattern of isograds it is only necessary that a_{H_2O} be a smoothly varying function of pressure and temperature. Thus metamorphic conditions will be restricted to a surface in P-T-a_{H_2O} space which will intersect the dehydration reaction surfaces. The resulting grid of intersection lines can be projected onto the P-T field. Grids of this type can be constructed from the observation of isograds in various metamorphic terranes (Hess, 1969; Carmichael, pers. comm., 1978). Although the functional relationship between a_{H_2O} and P-T might be expected to vary in different environments, this apparently does not lead to changes in topology of the grid. In fact, the topology appears to be insensitive to some fluctuations in a_{H_2O} and is commonly consistent with available experimental phase equilibria determined under the condition of $a_{H_2O} = 1$. However, a change in the functional relationship between a_{H_2O} and P-T might produce a noticeable shift of the projected dehydration reactions with respect to the stability field of the aluminum silicates.

The great number of dehydration reactions makes it necessary to select a few common ones to define boundaries of metamorphic grade, preferably in rocks of common composition and sensitive to metamorphic change. For this reason, Winkler (1976) chose, for the higher grades of metamorphism, reactions in pelitic rocks containing quartz, biotite, muscovite, and plagioclase.

Minerals in quartz-bearing pelitic rocks are represented approximately in the system Al_2O_3-K_2O-FeO-MgO. Their compositions may be projected either through muscovite (Thompson, 1957), or through $KAlSi_3O_8$ component in alkali feldspar or in melt, onto the plane Al_2O_3-FeO-MgO (AFM projection). In addition, mineral compositions have been projected through albite and anorthite components in plagioclase. Dehydration reactions which plot as a surface in P-T-a_{H_2O} space are represented by an exchange of tie lines on an AFM projection. In Figure 2, a portion of a reaction grid in pelitic rocks taken from Hess (1969) and Carmichael (pers. comm., 1978) is shown. The boundary between low grade and medium grade metamorphism is defined by five reactions which give rise to staurolite and/or cordierite in rocks consisting at lower grade of chlorite-muscovite-almandine. The boundary between medium grade and high grade metamorphism is defined by three reactions limiting the stability of muscovite in the presence of quartz, plagioclase, and biotite. At low pressures, muscovite and biotite react to form cordierite and almandine. At higher pressures, muscovite in the presence of plagioclase decomposes to alkali feldspar and aluminum silicate. This reaction, like the melting curve of granitic rocks, is somewhat dependent on the composition of plagioclase. Its intersection with the melting curve of granitic rocks generates a melting curve involving muscovite (Storre and Karotke, 1971).

Reactions in other rocks may be used as approximate indicators of facies boundaries, if it is known that they take place at similar conditions. Within each facies, other dehydration reactions may be used to define isograds, separating metamorphic zones. The sequence of isograds provides an approximate indication of pressure.

Table 1

Subdivisions of Metamorphic Grade

Turner and Verhoogen (1960) and Winkler (1967)	Turner 1968	Winkler (1976)	This paper
Granulite facies	Granulite facies	High grade; regional hypersthene zone	Granulite facies
Amphibolite facies; highest subfacies without muscovite	Amphibolite facies	High grade	Upper amphibolite facies
Amphibolite facies, all subfacies with muscovite	Amphibolite facies	Medium grade	Lower amphibolite facies
Greenschist facies	Greenschist facies	Low grade	Greenschist facies

On the metamorphic map (Fig. 3), the greenschist facies is undivided. Pelitic rocks typically contain chlorite and sericite, accompanied at higher grades by biotite and almandine. Basic rocks are characterized by the assemblage chlorite-epidote-actinolite in the lower greenschist facies and by the assemblage epidote-almandine-hornblende in the upper greenschist facies.

Three of the five reactions defining the lower boundary of the amphibolite facies in pelitic rocks extend to high pressures:

chlorite + almandine + muscovite
biotite + staurolite + quartz + H_2O (R.1)

chlorite + staurolite + muscovite + quartz
biotite + Al silicate + H_2O (R.2)

chlorite + muscovite + quartz
biotite + Al silicate + cordierite + H_2O (R.3)

At relatively high pressures, reaction (1) leads to the first appearance of staurolite in muscovite- and biotite-bearing rocks. Pelitic rocks of common composition will plot within the staurolite-almandine-biotite field and, consequently, reactions (2) and (3) will not be recorded in these rocks. However, with decreasing pressure, mineral compositions are shifted towards the F corner of the AFM diagram. Rocks of common composition will in this case plot to the right of the staurolite-biotite tie line and it becomes possible to recognize reactions (2) and (3). Thus, at relatively high pressures, the lower boundary of the amphibolite facies is characterized by the appearance of either staurolite, Al silicate, or cordierite-Al silicate in muscovite- and biotite-bearing rocks.

At relatively high pressures, the bulk composition of pelitic rocks commonly permits reaction (4) to occur. Thus staurolite disappears according to the reaction

staurolite + muscovite + quartz
biotite + Al silicate + almandine + H_2O (R.4)

In basic rocks, plagioclase (An 17) becomes stable at somewhat lower grade than given by reaction (1). Within the amphibolite facies, the anorthite content of plagioclase from amphibolites can be used as a measure of metamorphic grade (Wenk and Keller, 1969). In more detailed work, this method may prove useful. Epidote remains stable in amphibolites to about the boundary between lower and upper amphibolite facies. In calc-silicate rocks, diopside appears at about the greenschist facies – amphibolite facies boundary.

At intermediate pressures, the decomposition of muscovite in the presence of quartz and plagioclase marks the beginning of high grade metamorphism:

muscovite + plagioclase + quartz
K feldspar + Al silicate (R.5)

The composition of plagioclase will shift the position of this equilibrium. But for a constant plagioclase composition, reaction (5) is represented by a surface in P-T-a_{H_2O} space.

The intersection of reaction (5) with the melting curve of granitic rocks generates the reaction

muscovite + plagioclase + quartz + H_2O
melt + Al silicate (R.6)

According to Storre and Karotke (1971), this reaction (involving albite) has a steep negative slope.

In order to estimate prevailing pressures, it is significant to distinguish, whether reaction (5) or (6) has taken place, i.e. to distinguish whether the maximum stability of muscovite has been reached in unmigmatized gneisses or in migmatites.

Within the upper amphibolite facies the following reaction occurs:

biotite + Al silicate + plagioclase + quartz
cordierite + almandine + K feldspar + H_2O (R.7)

Again, its intersection with the melting curve gives rise to the reaction

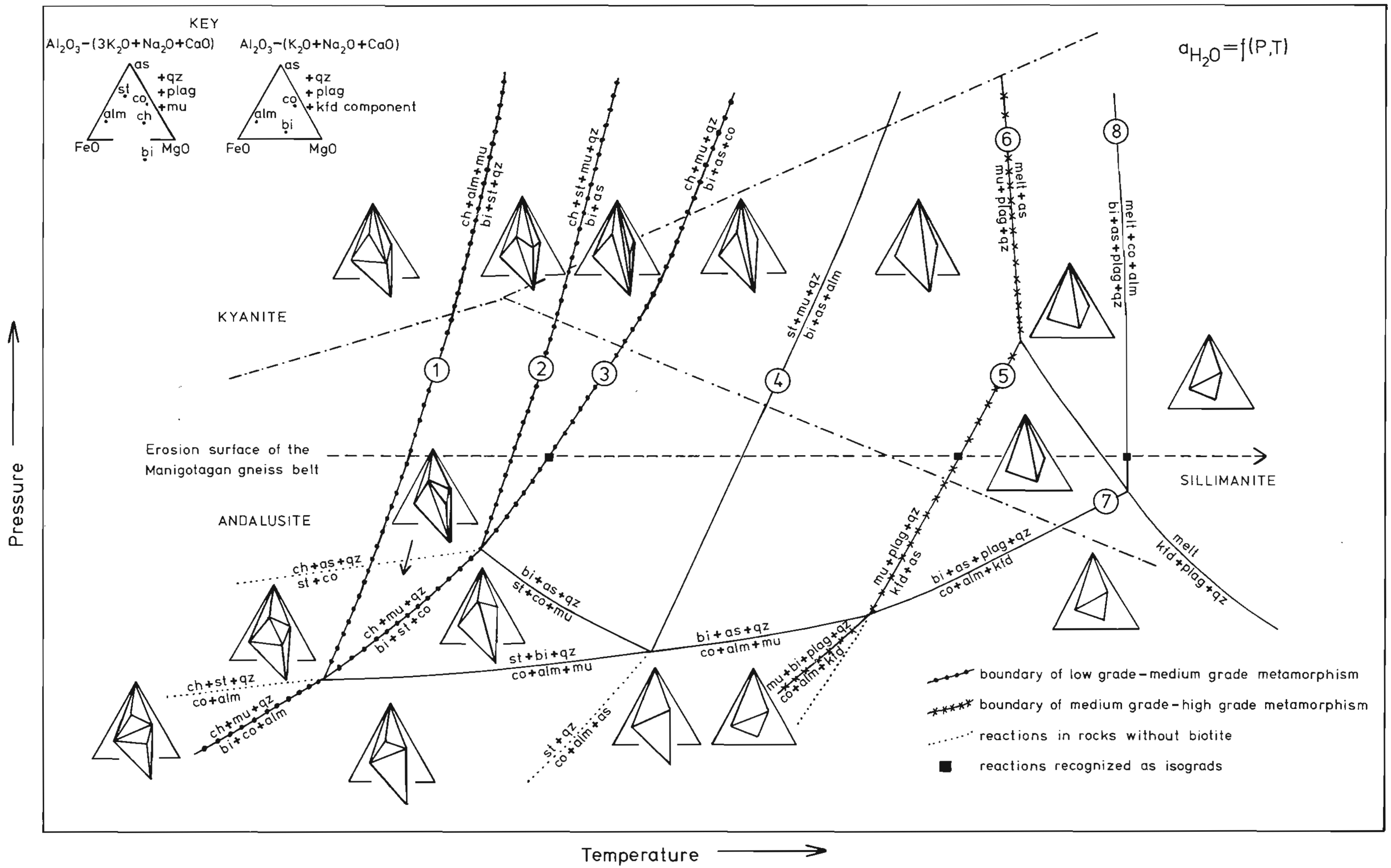
biotite + Al silicate + plagioclse + quartz + H_2O
melt + cordierite + almandine (R.8)

which is shown with a slope similar to that of reaction (6). The pressure range of regional metamorphism is such that most erosion surfaces will intersect reaction (8) rather than reaction (7).

The beginning of the granulite facies is defined by the appearance of orthopyroxene in pelitic and basic rocks, without regard to the specific reactions producing orthopyroxene (Winkler, 1976). Granulite terranes typically show no evidence of anatexis.

SACHIGO SUBPROVINCE

The Sachigo subprovince is characterized by volcanic belts which bifurcate to surround oval-shaped granodioritic batholiths. Some of these intrusions, particularly near their contacts, display a concordant foliation. Others are massive and appear to be unmetamorphosed. The supracrustal rocks in the western part of the Sachigo subprovince most commonly have mineral assemblages indicative of the lower amphibolite facies. In the centre of some of the larger belts (e.g. Oxford Lake), small areas of greenschist facies rocks are preserved and the metamorphic grade increases towards the margins of these belts (Weber and Scoates, 1978). Furthermore, Weber and Scoates (1978) describe a regional increase in metamorphic grade towards the northwest, culminating in granulite facies metamorphism in the Pikwitonei subprovince.



qz	—	quartz	st	—	staurolite
plag	—	plagioclase	co	—	cordierite
kfd	—	K feldspar	alm	—	almandine
mu	—	muscovite	ch	—	chlorite
as	—	aluminum silicate	bi	—	biotite

Figure 2. Reaction grid for pelitic rocks modified from Hess (1969) and Carmichael (pers. comm., 1978). Metamorphic reactions (R.1) to (R.8) are described in the text. Key to mineral abbreviations.

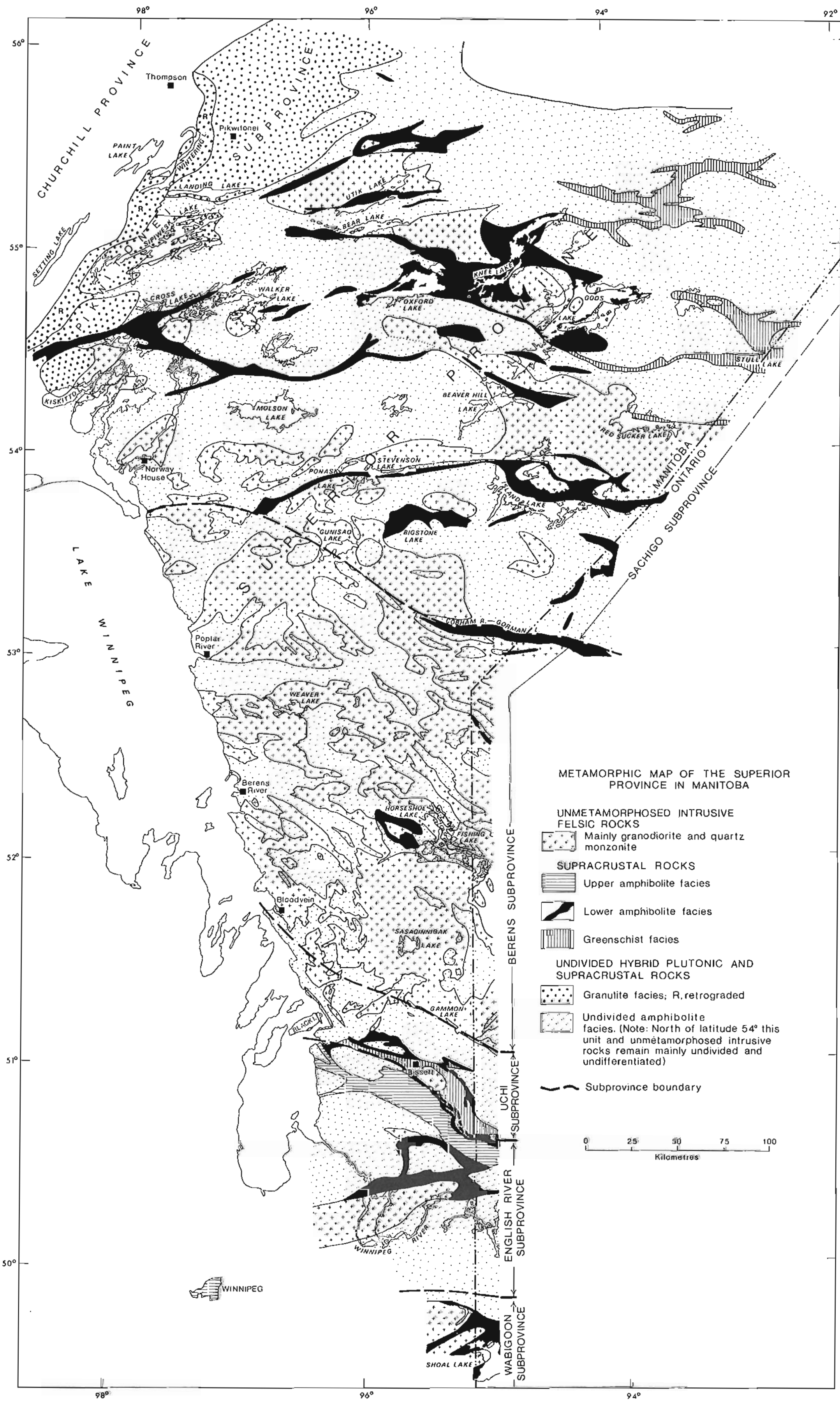


Figure 3. Metamorphic map of the Superior province in Manitoba.

In the eastern part of the Sachigo subprovince, rocks have been assigned to the greenschist facies on the basis of sparse petrographic information. The metamorphic grade of the gneissic granodioritic batholiths is difficult to establish. Their primary mineral assemblage may be taken to indicate partial melting, i.e. conditions of the upper amphibolite facies. However, their gneissic structure is taken as an indication that they are synorogenic intrusions and probably were subjected to the same metamorphic grade as the greenstone belts. This would not be reflected in mineralogical changes; accordingly, the metamorphic grade is shown as undivided amphibolite facies. The supracrustal rocks and the granodioritic batholiths have been retrograded locally to chlorite-bearing rocks indicative of the greenschist facies. The occurrence of cordierite and andalusite in metasedimentary rocks (Weber and Scoates, 1978; Herd and Ermanovics, 1976) indicates relatively low pressure metamorphism.

In the Oxford Lake-Knee Lake area, the volcanic Hayes River Group is overlain unconformably by the sedimentary Oxford Lake Group. Weber and Scoates (1978) recognized two metamorphic events, one predating and the other postdating the deposition of the Oxford Lake Group. A zircon U-Pb age of 2700 Ma (Weber and Scoates, 1978) from a metadacite of the Hayes River Group is considered to be a minimum age for the deposition of this group and gives an upper limit for the age of the earlier metamorphism. Two other unconformities occur in the Sachigo subprovince, one at Island Lake (Herd and Ermanovics, 1976), and the other at Cross Lake (Rousell, 1965). Apparently these unconformities represent only a short time interval because it is not possible to distinguish two periods of metamorphism. A zircon U-Pb age of 2700 Ma, obtained from the foliated granodiorite below the unconformity at Cross Lake (R.K. Wanless, pers. comm.), presumably dates the period of metamorphism in this area.

PIKWITONEI SUBPROVINCE

An area underlain by rocks of the granulite facies, west of the Sachigo subprovince, constitutes the Pikwitonei subprovince. A typical mineral assemblage, developed in rocks of appropriate composition, is plagioclase-clinopyroxene-orthopyroxene-garnet. Bell (1971) suggested that the Pikwitonei subprovince was separated from the Sachigo subprovince by an unconformity. An area of migmatite to the west of the Pikwitonei subprovince was assigned by Bell (1971) to the Wabowden subprovince of the Churchill province. Along the western edge of the Pikwitonei subprovince, the granulites have been retrograded to migmatites of the amphibolite facies. These rocks are similar to migmatites of the Wabowden subprovince, which might also represent retrograded Pikwitonei granulites. Recently Cranstone and Turek (1976) obtained an isochron of 2760 Ma (^{87}Rb decay constant = $1.42 \times 10^{-11} y^{-1}$) from Wabowden migmatitic gneisses. This was interpreted as the age of formation of the migmatitic gneisses from Pikwitonei rocks, which implies that the granulite metamorphism is older than 2760 Ma. On the other hand, Weber and Scoates (1978) report that the boundary between Pikwitonei granulites and rocks of the Sachigo subprovince, although marked by an increase of metamorphic grade, is essentially gradational, i.e. the boundary is an isograd and not an unconformity. This observation prompted them to interpret the 2760 Ma age as a minimum age of either the granulite metamorphism or of the original rocks. Accordingly, 2760 Ma would be an upper limit for both periods of granulite metamorphism (M_1 and M_2) recognized by Hubregtse (1977). He also reported kyanite in a plagioclase-quartz mobilizate developed during the second period of granulite metamorphism. Because kyanite can be formed in veins, its occurrence in a mobilizate does not necessarily indicate high pressures.

BERENS SUBPROVINCE

South of the Sachigo subprovince and in places separated from it by zones of cataclastic rocks, is the Berens subprovince. The boundary is marked by steep aeromagnetic gradients and coincides approximately with the -30 mgal Bouguer gravity anomaly. The subprovince, characterized by a dearth of supracrustal rocks, consists mainly of gneissic batholiths and unmetamorphosed granitic batholiths. The gneissic batholiths include a few remnants of supracrustal rocks. Larger areas of supracrustal rocks occur only at Horseshoe Lake and Gorman Lake. The coexistence of quartz, muscovite, and sillimanite in these rocks indicates a metamorphic grade corresponding to the lower amphibolite facies. The metamorphic grade of the gneissic batholiths is shown as undivided amphibolite. Locally these rocks and some interlayered amphibolites have been retrograded to the greenschist facies.

UCHI SUBPROVINCE

The boundary between the Berens subprovince and the Uchi subprovince in Manitoba is difficult to define. Following Douglas (1973), the volcanic rocks at Bissett as well as the plutonic rocks, paragneisses, and amphibolites north of Bissett have been included in the Uchi subprovince. Steep aeromagnetic gradients support the choice of this boundary. The metamorphic grade ranges from greenschist facies in volcanic rocks at Bissett to lower amphibolite facies in the paragneisses nearer to the Berens subprovince. At Gammon Lake the assemblage biotite-sillimanite-cordierite was observed (Ermanovics, 1970a). Contact metamorphism associated with some plutonic rocks (McRitchie and Weber, 1971b) has not been shown in Figure 3. There is pronounced retrograde metamorphism along the Wanipigow River lineament extending from Black Island to Bissett. Turek (1971) recognized three groups of age determinations. The two oldest events at about 2700 Ma and 2500 Ma were interpreted to date periods of intrusion and metamorphism. The third event represents a period of cataclasis. Krogh et al. (1974) report two groups of age determinations at 2700 Ma and 2900 Ma, presumably also representing periods of metamorphism. The 2900 Ma age might reflect an older basement. The metamorphic map (Fig. 3) shows the main regional zonation (M_{1A}) recognized by McRitchie and Weber (1971b). This might be related to the 2700 Ma event (Turek, 1971). The main zonation is separated from a weak secondary regional zonation (M_2) by a time interval marked by an unconformity below the youngest supracrustal rocks (San Antonio Formation).

ENGLISH RIVER SUBPROVINCE

The supracrustal rocks of the Uchi subprovince grade into paragneisses of the Manigotagan gneiss belt south of Bissett, which forms part of the English River subprovince. The transition is marked by a lithologic change from volcanic to sedimentary rocks and by an increase in metamorphic grade. Mapping by McRitchie (1971) indicated a distinct plutonic domain to the south of the Manigotagan gneiss belt. Recent work by the Ontario Division of Mines, referred to in the paper by Thurston and Breaks (1978), indicated that the whole English River subprovince consists of a northern supracrustal domain and a southern plutonic domain. These divisions have been designated by Beakhouse (1977) as the Ear Falls-Manigotagan gneiss belt and the Winnipeg River batholithic belt.

The metamorphic history of the Manigotagan gneiss belt is well documented (McRitchie and Weber, 1971b). Disregarding a weak secondary zonation (M_2), reactions indicated by the main regional metamorphic zonation (M_{1A}) on the present erosion surface are shown in Figure 2. The sequence of isograds indicates low pressure metamorphism. The lower boundary of the amphibolite facies is typically defined by reaction (3). Reaction (5) takes place at slightly higher grade than the andalusite-sillimanite transition. Within the range of the upper amphibolite facies, an isograde based on reaction (8) more or less coincides with the beginning of anatexis. The zonation in the Manigotagan gneiss belt may be contrasted with the sequence of isograds in the Flin Flon-Snow Lake belt recording reactions (1), (2), (4), (6), and (8) (Bailes and McRitchie, 1978).

A similar zonation marks the transition from the Manigotagan gneiss belt to the Bird River greenstone belt, 25 km north of the Winnipeg River, which occurs near the boundary of the two divisions of the English River subprovince (McRitchie, 1971; Trueman et al., 1975). The metamorphic grade of the Winnipeg River batholithic belt is shown as undivided amphibolite facies, similar to that of the Berens subprovince.

WABIGOON SUBPROVINCE

The boundary between the English River subprovince and the Wabigoon subprovince is poorly exposed and hence not well defined. It is marked, by a transition to a greater abundance of volcanic rocks and in places it may be represented by faults or intrusive contacts (Beakhouse, 1977). The supracrustal rocks of the Wabigoon subprovince in Manitoba have been metamorphosed in the lower amphibolite facies (Janes and Malyon, 1977).

CONCLUSIONS

The Sachigo, Berens, Uchi, and Wabigoon subprovinces share geological and structural similarities. The Sachigo, Uchi, and Wabigoon subprovinces are characterized by belts of supracrustal rocks which surround oval-shaped batholiths. This structure is reflected in a distinct aeromagnetic pattern (Wilson, 1971) which is also present in the Berens subprovince. The rarity of supracrustal rocks in the Berens subprovince suggests that it is an uplifted terrane of the type now exposed in the Sachigo subprovince. The characteristic structure of the volcanic belts has been attributed by Brown (1973) to vertical rather than horizontal movements. The degree of deformation of the volcanic rocks is greatest near their contacts; subsequently, primary structures are better preserved in the thicker portions of rocks at the triple junctions of bifurcating belts. The grade of metamorphism ranges from greenschist facies to lower amphibolite facies.

The English River subprovince is characterized by metasedimentary rocks which produce a distinct linear aeromagnetic pattern. A well-developed metamorphic zonation marks the boundary between the Uchi subprovince and the English River subprovince. To the south, the grade of metamorphism decreases from the Manigotagan gneiss belt to the Bird River greenstone belt which is situated along the proposed boundary of a two-fold division of the English River subprovince. The northern part of the English River subprovince (Ear Falls-Manigotagan gneiss belt) as well as adjacent volcanic areas, the Uchi subprovince to the north and the Bird River greenstone belt to the south, are underlain by a thick upper crust and a thin total crust (Hall, 1971; Beakhouse, 1977). Possibly this permitted a greater heat flow, giving rise to a belt of high grade metamorphism flanked by lower grade metamorphic rocks.

The presence of cordierite and andalusite in most subprovinces of the Superior province in Manitoba indicates low pressure metamorphism. The absence of pelitic metasedimentary rocks in the Pikwitonei subprovince makes an estimation of metamorphic pressure in that subprovince uncertain.

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METAMORPHISM IN THE SUPERIOR PROVINCE OF NORTHWESTERN ONTARIO AND ITS RELATIONSHIP TO CRUSTAL DEVELOPMENT

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Abstract

The early Precambrian Superior Province in northwestern Ontario comprises seven easterly-trending subprovinces that differ in lithology, structure, and metamorphism. Each subprovince is a primary lithologic sequence, and variation in lithology and metamorphism among subprovinces may reflect different stages in crustal evolution. Because of higher geothermal gradients metamorphic reactions occurred at higher crustal levels than in comparable Phanerozoic sequences. Consequently early Precambrian metamorphic assemblages cannot be used to estimate burial depths by comparison with Phanerozoic sequences.

The Wawa, Wabigoon, Uchi, and Sachigo subprovinces comprise isoclinally folded, metavolcanic-metasedimentary sequences (greenstone belts) intruded by large composite, weakly recrystallized granitic batholiths. These subprovinces represent deformed linear volcanic island chains. The larger greenstone belts are metamorphically zoned, with a greenschist facies core successively surrounded by a relatively narrow amphibolite facies zone and a discontinuous hornblende hornfels facies zone; subgreenschist assemblages are rare. The highest metamorphic grades and steepest metamorphic gradients are adjacent to the granitic batholiths which were the heat source for much of the metamorphism. Consequently the metamorphism reflects a contact rather than a regional event, although the inner parts of the greenschist facies zone may be the result of regional burial metamorphism. The relatively low metamorphic grade in the centre of the greenstone belts and the high thermal gradients at the margins of the belts indicate relatively shallow burial, probably less than 10 km. The higher grade metamorphic zones vary in width among belts and an increase in the width of the zones apparently reflects greater burial depths.

The Berens River subprovince contains more granitic plutons and relatively few greenstone belts which are entirely amphibolite facies. It represents a deeper more plutonic crustal level.

The Quetico and northern English River subprovinces are mainly migmatitic metaturbidites of middle to upper amphibolite and locally granulite facies. The southern English River subprovince is mainly orthogneiss, that may be older sialic basement, and variably recrystallized granitic plutons. These subprovinces represent deformed linear turbidite basins that developed between the volcanic island chains, but now show a deeper crustal level than the greenstone belt subprovinces.

Résumé

Dans le secteur nord-ouest de l'Ontario, la province du lac Supérieur, d'âge précambrien inférieur, comprend sept sous-provinces orientées vers l'est, qui diffèrent du point de vue de la lithologie, de la structure, et de l'intensité du métamorphisme. Chaque sous-province représente une succession lithologique primaire, et les variations de la lithologie et du métamorphisme d'une sous-province à l'autre reflètent probablement diverses phases d'évolution de la croûte terrestre. En raison de l'existence d'un gradient géothermique plus élevé à cette époque, les réactions métamorphiques se sont produites à moins grande profondeur, que dans les successions comparables du Phanérozoïque. Il n'est donc pas possible de comparer les successions phanérozoïques aux assemblages métamorphiques du Précambrien inférieur pour estimer la profondeur d'enfouissement pendant le Précambrien.

Les sous-provinces de Wawa, Wabigoon, Uchi et Sachigo comprennent des successions métavolcaniques et métasédimentaires caractérisées par des plis isoclinaux (zones de roches vertes), traversées par de grands batholithes granitiques, de caractère composite, faiblement recristallisés. Ces sous-provinces représentent des chaînes volcaniques insulaires, d'allure linéaire, qui ont été déformées. Les zones de roches vertes les plus vastes manifestent une zonalité métamorphique; leur centre, métamorphisé dans le faciès des roches vertes, est successivement entouré par une zone relativement étroite du faciès amphibolite, puis une zone discontinue du faciès des cornéennes à hornblende; les assemblages des sous-schistes verts sont relativement rares. Le métamorphisme le plus intense et les gradients de métamorphisme les plus prononcés sont observés à proximité des batholithes granitiques, qui ont été la principale source thermique et de métamorphisme. Ainsi, il s'agit davantage d'un métamorphisme de contact que d'un métamorphisme regional, bien que le noyau de la zone métamorphique du faciès schistes verts doit sans doute son caractère à un métamorphisme régional d'enfouissement. L'intensité relativement faible du métamorphisme au centre des zones de roches vertes, et le gradient thermique élevé sur les bords de ces zones indiquent que la profondeur d'enfouissement était relativement faible, probablement inférieure à 10 kilomètres. La largeur des secteurs plus fortement métamorphisés varie d'une zone à l'autre, et apparemment, toute augmentation de largeur reflète un accroissement de la profondeur d'enfouissement.

La sous-province de Berens River contient davantage de plutons granitiques, et relativement peu de zones de roches vertes qui appartiennent entièrement au faciès amphibolite. Elle représente un niveau crustal plus profond, de caractère davantage plutonique.

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Les sous-provinces de Quetico et du nord de English River sont surtout des métaturbidites de caractère migmatitique métamorphisées dans le faciès amphibolite moyenne à supérieure et parfois dans le faciès granulite. Le sud de la sous-province de English River est surtout constitué d'orthogneiss, qui représente peut-être l'ancien soubassement sialique, et de plutons granitiques plus ou moins recristallisés. Ces sous-provinces représentent des bassins à turbidites, déformés, d'allure linéaire, qui se sont constitués entre des chaînes volcaniques insulaires, mais qui correspondent maintenant à un niveau crustal plus profond que les sous-provinces de la zone des roches vertes.

INTRODUCTION

In the early Precambrian Superior Province of northwestern Ontario, west of 86°W, seven major easterly-trending subprovinces are delineated (Fig. 1); from south to north these are the Wawa, Quetico, Wabigoon, English River, Uchi, Berens River, and Sachigo. These subprovinces vary in width from 40 to 290 km and each has characteristic lithologies, structures, and metamorphic features. Lithologic boundaries between subprovinces are relatively sharp, and in many places are major faults; metamorphic boundaries are generally gradational. Each subprovince appears to comprise a primary lithologic sequence, the geographic disposition of which reflects the progressive development of the Superior Province. Consequently, variations in lithology, structure, and metamorphism should reflect different stages in the evolution of the early crust. Most of the boundary faults are probably later structures that emphasized the primary lithologic differences. This paper briefly reviews metamorphism in the various subprovinces, and emphasizes broad scale features rather than detailed metamorphic reactions.

The seven subprovinces can be broadly grouped into three major categories. The Wawa, Wabigoon, Uchi, and Sachigo subprovinces[1] comprise isoclinally folded metavolcanic-metasedimentary sequences (greenstone belts) intruded by large composite granitic batholiths that predominate areally over the metavolcanic-metasedimentary sequences. In the greenstone belts, metamorphic grade ranges from low greenschist and locally subgreenschist facies to mid-amphibolite facies; the highest grades are adjacent to the batholiths (Ayres, 1972; Ayres et al., 1972b).

In the Quetico and English River subprovinces, metamorphic grade ranges from amphibolite to local granulite facies. The Quetico subprovince appears to be mainly metasedimentary gneiss that was derived from a turbidite sequence and contains a relatively small proportion of granitic plutons (Fig. 1). The English River subprovince (Breaks et al., 1974; Breaks and Bond, 1977; Harris and Goodwin, 1976; Beakhouse, 1974a, 1977) consists of the northern Ear Falls-Manigotagan portion which is broadly similar to the Quetico subprovince, and a distinctly different southern Winnipeg River portion. The latter, which seems to be restricted to the area west of 89°W (Fig. 1), is mainly quartz dioritic to trondhjemitic orthogneiss intruded by less deformed granodiorite to granite plutons (Harris and Goodwin, 1976; Beakhouse, 1977; Breaks and Bond, 1977). The orthogneiss may be in part sialic basement upon which the northern metasedimentary sequence was deposited (Harris and Goodwin, 1976).

The Berens River subprovince (or Berens batholithic belt) is composed mainly of granitic plutons with less than 10 per cent recognizable metavolcanic-metasedimentary remnants at middle to upper amphibolite facies metamorphic grade. Ermanovics and Davison (1976) proposed that the Berens River subprovince in Manitoba represents a remobilized and uplifted equivalent of the Sachigo greenstone-granodiorite subprovince. Ayres et al. (1973), on the other hand, proposed that the Berens River subprovince in Ontario was a zone of deeper and more intense plutonic activity than either of the adjacent greenstone-granodiorite subprovinces. There is no evidence of major remobilization.

WAWA, WABIGOON, UCHI, AND SACHIGO SUBPROVINCES

In these four subprovinces, isoclinally folded, metavolcanic-metasedimentary sequences form linear, generally easterly trending greenstone belts that were intruded by composite batholiths ranging in composition from diorite to granite; the average composition is granodiorite. Associated with the batholiths are variable amounts of felsic gneiss of uncertain origin, some of which are remnants of pre-volcanic sialic crust (Ayres, 1974a, 1974b; Hillary, 1976).

Greenstone Belts

Origin of Subprovinces

In all the greenstone belts, the stratigraphic sequence is incomplete; tops are eroded and the bases are either not exposed, or intruded by granitic batholiths. The relatively small greenstone belts in northwestern Ontario (Fig. 1) appear to be isolated remnants of metavolcanic-metasedimentary sequences that once covered most of the greenstone-granodiorite subprovinces. The preserved sequences are up to 20 km thick and consist of subaqueous basaltic shield volcanoes 10-15 km thick capped by complex subaerial to subaqueous, andesitic to rhyolitic stratovolcanoes.

The lateral dimensions of individual volcanoes can only be crudely estimated, but the volcanoes appear to have been considerably larger than most greenstone belts but smaller than the subprovinces. For example, the Wabigoon subprovince, which is readily defined by the bordering gneissic and metasedimentary Quetico and English River subprovinces, ranges in width from 100 to 175 km (Fig. 1) and has an exposed length of 800 km. This subprovince was probably originally a linear, volcanic island chain comprising a number of individual volcanoes whose lower and middle parts coalesced to form a continuous unit. It is now highly deformed and dismembered.

The south margin of the Wabigoon subprovince is largely a fault boundary, but locally metasediments of the Quetico subprovince appear to be facies equivalents of the volcanic rocks (Ayres, 1969a; Mackasey et al., 1974). Elsewhere the abrupt change across the fault from metavolcanics to metasediments supports a primary facies origin for the subprovince boundary.

Along the north margin, the Wabigoon subprovince is mainly in contact with the Winnipeg River portion of the English River subprovince (Fig. 1). According to Breaks and Bond (1977) the subprovince boundary is not faulted; however, the nature of the boundary is not known because of poor exposure and the presence of younger granitic sills. If the

[1] The existence of the subprovinces was recognized more or less simultaneously by several workers (Wilson, Goodwin, Stockwell), who applied different names. In some cases, these names were used informally for several years prior to formal publication, and as a result there is a conflict in nomenclature. The terminology of Goodwin (1966, 1968) appears to have priority of publication, but most workers use the terminology of Stockwell (in Wanless et al., 1968) and Douglas (1974), although the terminology of Wilson (1971) is used by many workers in the western Superior Province. The terminology used herein is that of Douglas (1974), although the positions of the subprovince boundaries do not completely agree with those of Douglas (1974) or of other workers. Boundaries have been modified from Ayres et al. (1972a).

Winnipeg River portion is in part older sialic basement (Harris and Goodwin, 1976), then the Wabigoon-English River subprovince boundary may be in part an unconformity modified by younger intrusions, some of which may have been intruded along faults. The boundary is thus a primary lithologic contact.

In the northern Ear Falls-Manigotagan portion of the English River subprovince the metasediments are largely of volcanic origin and may be facies equivalents of metavolcanic rocks of the Uchi subprovince to the north (van de Kamp, 1973; Beakhouse, 1974b; Breaks and Bond, 1977), a relationship similar to that at the Wabigoon-Quetico subprovince boundary. The Ear Falls-Manigotagan metasediments may unconformably overlie the Winnipeg River orthogneiss (Harris and Goodwin, 1976), and thus represent a sedimentary basin between the Uchi and Wabigoon subprovinces.

A primary origin for these subprovinces is supported by the parallel linearity of subprovinces, the presence of abrupt lithologic changes at the boundaries, the lithologies present within the subprovinces, and the facies transitions at some boundaries. Thus the Wabigoon subprovince probably represents a former linear chain of volcanoes that developed along a fundamental fracture system in the early Precambrian crust. The chain was bounded on the south by a turbidite sedimentary basin now represented by the Quetico subprovince, and on the north it overlapped unconformably onto Winnipeg River sialic basement of the English River subprovince. A second turbidite sedimentary basin developed in the northern part of the English River subprovince. The volcanic chain is now highly modified by deformation and batholith emplacement.

The other greenstone-granodiorite subprovinces may also represent linear volcanic chains of similar or different ages, but this cannot be documented at present. However, the similarity of rock units with the Wabigoon subprovince supports a similar origin.

Geothermal Gradient

Any discussion of early Precambrian volcanism and metamorphism must consider the higher geothermal gradient and the correspondingly less stable crustal conditions prevalent at that time. In the greenstone-granodiorite subprovinces a higher gradient resulted from a greater abundance of radioactive heat-producing elements (Lambert, 1976), and increased heat flow because of upward movement of magma. The former affected all parts of the early Precambrian crust, but the latter was concentrated in the volcanic chains. The higher heat flow and active magma generation at depth may have been major factors in the localization of granitic plutons. These plutons are generally more abundant in the greenstone-granodiorite subprovinces than in the metasedimentary portions of the Quetico and English River subprovinces.

In the upper 40 to 50 km of the crust and mantle, estimated Precambrian gradients based on heat productivity are higher than present gradients by a factor of at least two, but at greater depths the difference is less (Green, 1975; Hargraves, 1976; Lambert, 1976). In areas of active volcanism and plutonism however, geothermal gradients were probably even higher because ascent of magma transferred heat upward.

The inference to be drawn from such high geothermal gradients is that low pressure facies series should prevail. Furthermore, metamorphic grade should be relatively high in the greenstone belts and low in the English River and Quetico subprovinces. However, the observed metamorphic grade is actually higher in the Quetico and English River subprovinces than in the greenstone belts, and there is an inverse relationship between metamorphic grade and magmatism. This suggests that the Quetico and English River subprovinces represent deeper crustal levels than do the greenstone-granodiorite subprovinces.

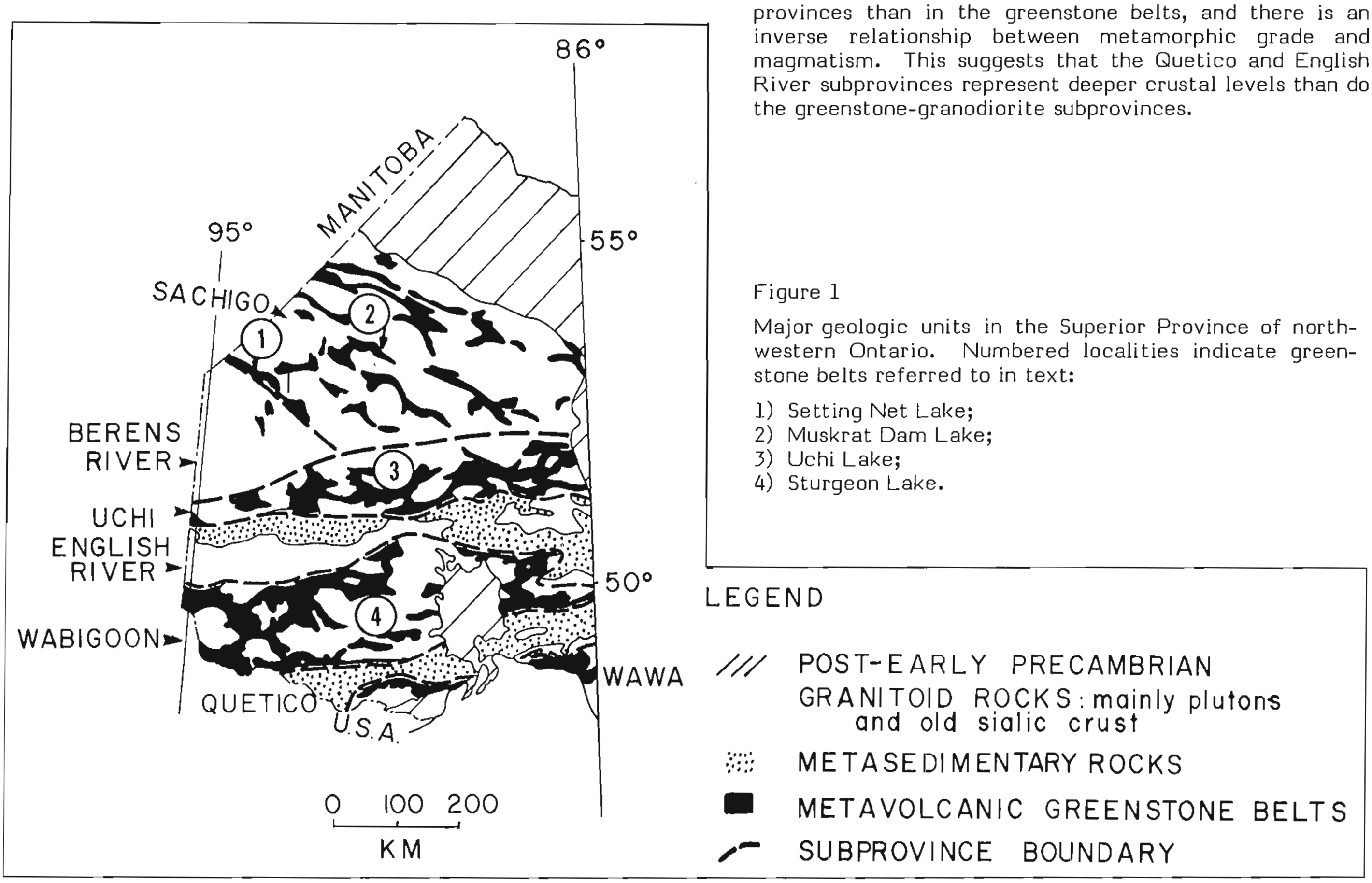

Figure 1

Major geologic units in the Superior Province of northwestern Ontario. Numbered localities indicate greenstone belts referred to in text:

1) Setting Net Lake;
2) Muskrat Dam Lake;
3) Uchi Lake;
4) Sturgeon Lake.

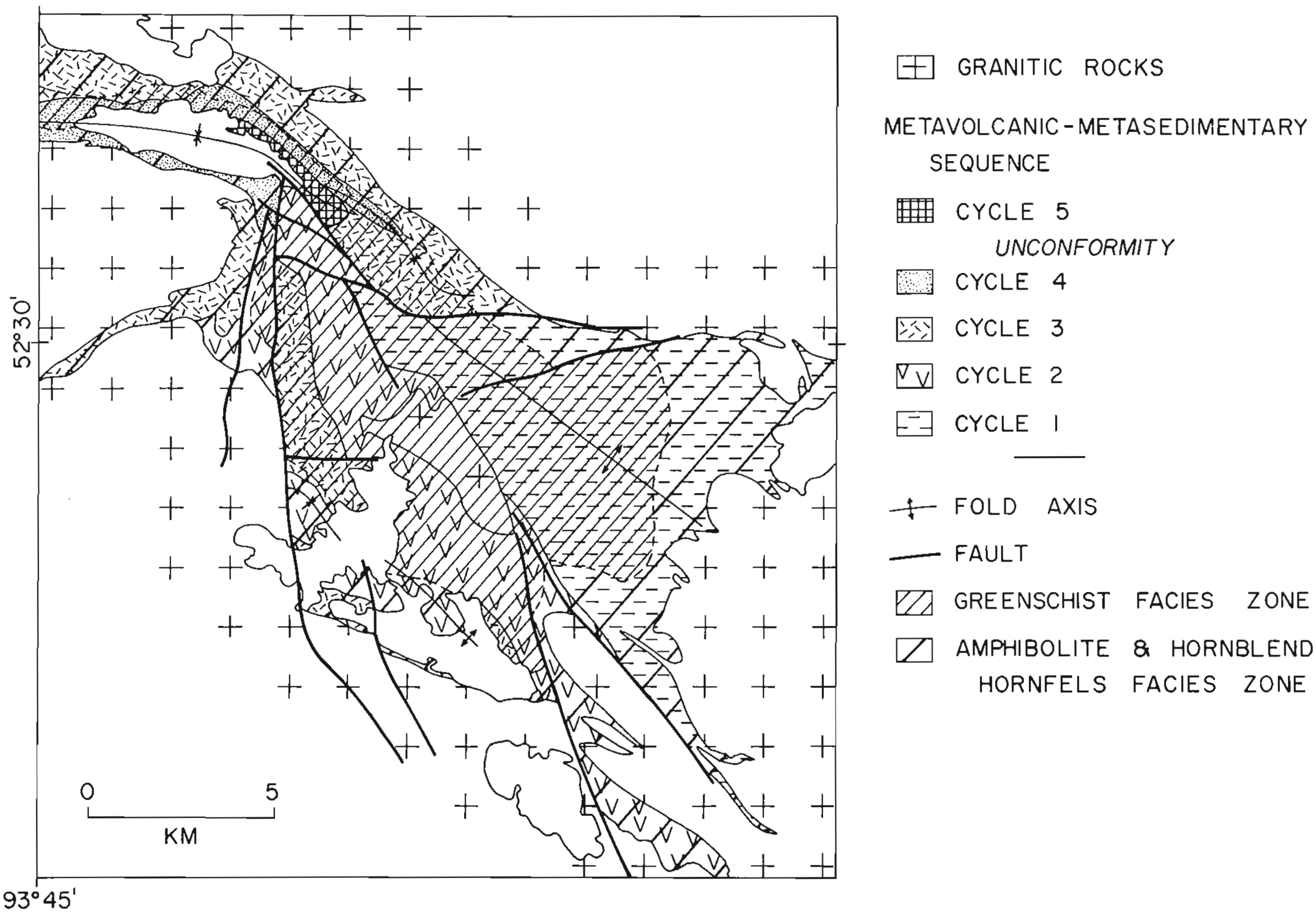

Figure 2. Metamorphic zones and simplified stratigraphy of the Setting Net Lake greenstone belt. The hornblende hornfels zone is not shown separately on this figure and on Figure 3 because it is discontinuous and narrow.

Metamorphism

Only two greenstone belts in northwestern Ontario have been studied in detail: Muskrat Dam Lake (Ayres, 1969b) and Setting Net Lake (Ayres, 1974a) (Figs. 1,2,3). In these and most other greenstone belts an inner greenschist facies zone is surrounded by an outer amphibolite facies zone and a locally developed marginal hornblende hornfels facies zone adjacent to the granitic batholiths (Ayres, 1972; Ayres et al., 1972b). Zone boundaries are gradational and difficult to define because diagnostic index minerals are absent.

A greenschist facies zone is present in most belts wider than 5 km, but is generally absent in narrower belts. In the narrower belts metamorphic grade is entirely amphibolite and hornblende hornfels facies. In the wider belts the amphibolite and hornblende hornfels facies zones vary in width both within belts and among belts. At Setting Lake (Fig. 2), the hornfels zone is up to 300 m wide; the amphibolite facies zone ranges in width from 400 to 3000 m, and averages about 1500 m. At Muskrat Dam Lake both zones are somewhat wider. The hornfels zone is as wide as 1500 m, and the amphibolite facies zone ranges in width from 400 to 5000 m (Fig. 3). The amphibolite facies zone increases in width eastward along the belt averaging 1500 m in the west and 3000 m in the east. The greenschist facies zone has a maximum width of 9 km at Setting Net Lake and 15 km at Muskrat Dam Lake.

In some belts this simple zonal pattern is distorted by albite-epidote, hornblende, and locally pyroxene hornfels facies contact aureoles developed around small plutons. Most of these aureoles are relatively narrow and are characterized by low pressure facies series assemblages. Some of the older synvolcanic aureoles were retrograded during regional metamorphism, whereas the younger postkinematic aureoles are superimposed on the regional metamorphism.

In the greenstone belts the predominant rock types are mafic to felsic flows and fragmental rocks, greywacke, and siltstone. In the dominant mafic to intermediate metavolcanic rocks the characteristic greenschist facies mineral assemblage is chlorite and/or actinolite – albite – epidote ± carbonate, and in the less common felsic metavolcanics and metasediments it is biotite and/or chlorite – albite – epidote – quartz ± muscovite ± carbonate. Detailed assemblages are listed in Ayres (1969b, 1974a). In the greenschist facies zone of the wider greenstone belts, there is probably a progressive decrease in metamorphic intensity inward, but definitive data for such a change are sparse (Thurston and Breaks, 1978). Prograde subgreenschist facies assemblages have been reported only in the Uchi Lake area (Fig. 1; Thurston and Breaks, 1978), but retrograde zeolites and prehnite are common. These minerals occur in microscopic veins in both greenschist and higher grade units, and locally replace plagioclase in the hornblende hornfels facies zone.

Unlike the Abitibi Belt of northeastern Ontario (Jolly, 1974), there is no regional pattern of chlorite, actinolite, and biotite distribution. Chlorite is present throughout the greenschist facies zone and appears to be related more to P_{CO_2} than to temperature or load pressure. Chlorite is commonly associated with carbonate in both mafic and felsic

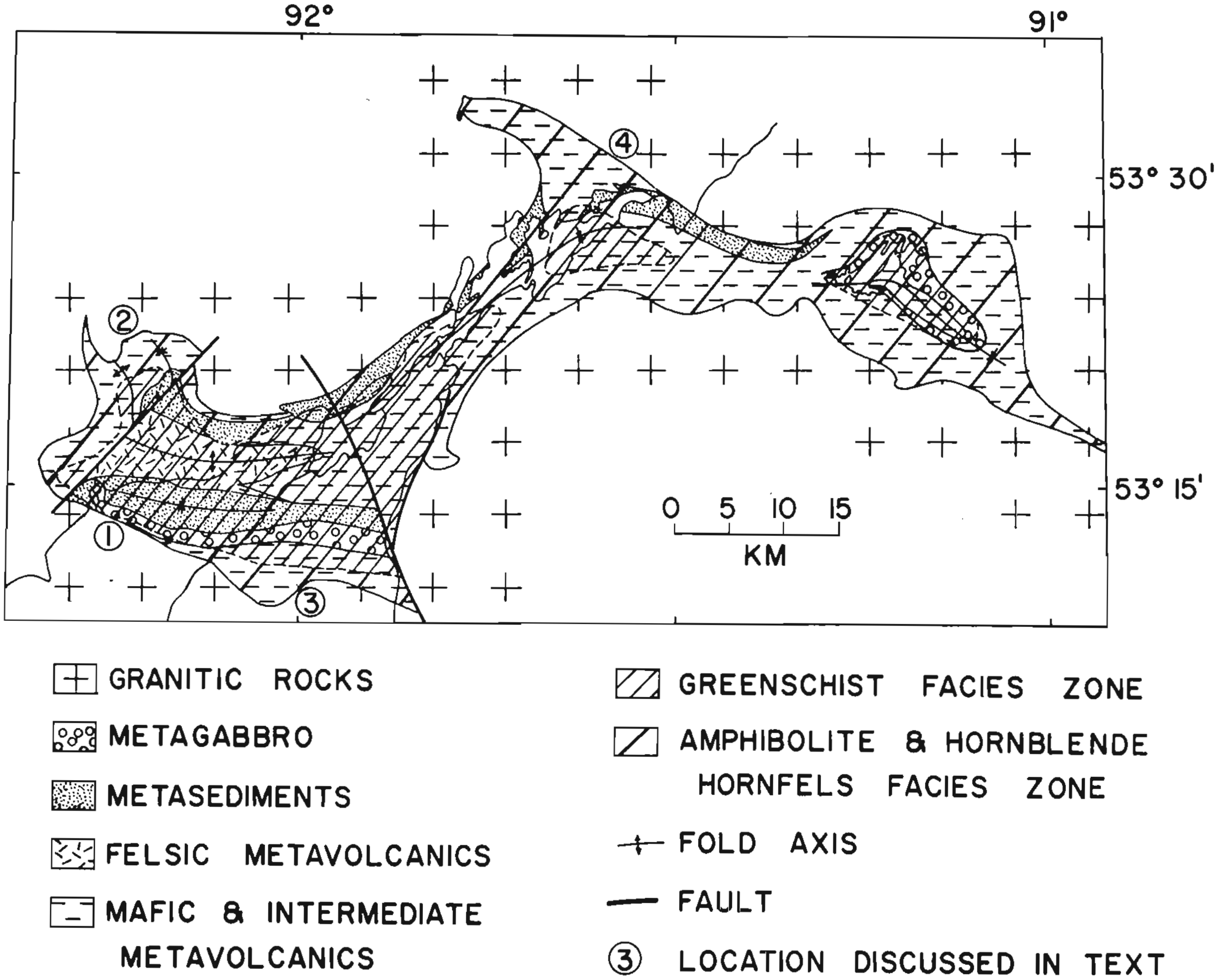

Figure 3. Metamorphic zones and simplified lithology of the Muskrat Dam Lake greenstone belt (Ayres, 1969b).

rocks, but in carbonate-poor rocks of similar grade the dominant mafic mineral in mafic rocks is actinolite and in felsic rocks, biotite.

Primary textures and minerals are best preserved in actinolite- and biotite-bearing assemblages, and are largely destroyed in chlorite-bearing assemblages. This association suggests that biotite- and actinolite-bearing assemblages did not develop from earlier chlorite-bearing assemblages. The transition from chlorite-bearing assemblages to biotite- and actinolite-bearing assemblages is more a function of P_{CO_2} than temperature. High P_{CO_2} apparently extended the stability field of chlorite to higher temperatures and facilitated the breakdown of primary minerals, particularly plagioclase. Thus, in areas of variable P_{CO_2}, as indicated by variable development of carbonate, the distribution of chlorite, actinolite, and biotite may not be reliable indicators of metamorphic grade, particularly in relatively small greenstone belts.

In the amphibolite facies most mineral assemblages reflect low to intermediate pressure facies conditions. The characteristic assemblage in mafic and intermediate metavolcanic rocks is hornblende – plagioclase, and in felsic metavolcanics and metasediments, plagioclase – quartz – biotite ± garnet ± muscovite. In the relatively rare pelites the following minerals and mineral combinations are locally developed: staurolite, staurolite – andalusite, staurolite – andalusite – cordierite, andalusite – cordierite, and cordierite. Sillimanite is rare except in greenstone belts near the boundaries with the Quetico, English River, and Berens River subprovinces (Fig. 4; see e.g. Pryslak, 1976; Wood, 1975). Only four occurrences of kyanite are known (Fig. 4). Three of these are in amphibolite facies terrane near subprovince boundaries, where the development of kyanite may reflect local high pressures related to deformation. The fourth occurrence is in the ore zone and adjacent footwall rocks of the Mattabi deposit in the Sturgeon Lake greenstone belt of the Wabigoon subprovince (Fig. 1) (Franklin et al., 1975; Trowell, 1974). This kyanite may reflect abnormal bulk compositions rather than high pressure metamorphism. A similar kyanite occurrence has been reported by Tolman (1951) from the Normetal ore deposit in Quebec.

A hornblende hornfels facies zone has been recognized only in mafic metavolcanic rocks. Except for the local development of clinopyroxene and garnet in the outermost part of the hornfels zone, the amphibolite facies and hornfels facies zones are mineralogically so similar that they can be distinguished only on textural evidence; the amphibolite is well foliated, whereas the hornfels is granoblastic and massive to gneissic.

The boundary between the greenschist and amphibolite facies is best defined by the transition from albite to oligoclase-andesine, and by systematic changes in the amphiboles. In most of the greenschist facies, the amphibole is colourless to pale green and fibrous, whereas in the uppermost part of the greenschist facies, the amphibolite facies, and the hornblende hornfels facies zones the colour of the amphibole progressively deepens to a dark green tint and grains become more equant. These changes may correspond to the transition from actinolite to hornblende, but they do not coincide with the greenschist-amphibolite facies boundary defined from plagioclase composition.

Metamorphic Zoning

The metamorphic zoning transects both stratigraphic and structural trends and must be a relatively late feature. The increase in metamorphic grade toward the granitic batholiths suggests that the metamorphism is related to batholith emplacement. Metamorphic zoning apparently developed in response to horizontal temperature gradients that were highest at the edges of the greenstone belts, adjacent to the magmatic heat source, and decreased inward (Fig. 5). The gradients differ from belt to belt, a feature reflected in the variable widths of the combined amphibolite facies and hornblende hornfels facies zones.

The relationship between metamorphic zoning and stratigraphy is well illustrated in the Setting Net Lake area (Fig. 2). Here a metavolcanic-metasedimentary sequence 7.5 km thick is composed of five groups or cycles that are progressively younger to the northwest, along the axis of the belt. The configuration of the belt is controlled by isoclinal folding and by granitic batholiths. The metamorphic zoning shows a crude relationship to the stratigraphy, in that the amphibolite plus hornblende hornfels facies zone is widest in cycle 1, suggesting that the lower part of the volcanic sequence is somewhat more metamorphosed than the upper part. However, stratigraphy is not the main control on metamorphic zone development because the higher grade zones transect the four lower cycles.

Similarly structure was not a major control of metamorphic zone development. In most greenstone belts, major isoclinal folds parallel belt boundaries, and metamorphic zones are subparallel to fold axes. However, where folds are transected by belt boundaries, as at the ends of many greenstone belts, metamorphic zones cut across folds (Figs. 2,3) and apparently developed after the folds. Furthermore, in the northwestern part of the Muskrat Dam Lake belt (locality 2, Fig. 3), gneissosity related to hornblende hornfels facies metamorphism transects and obliterates the axial planar foliation of isoclinal folds (Ayres, 1969b). Although folding appears to predate the major metamorphic event, many faults postdate the metamorphism, and metamorphic zones are offset along faults.

The increase in metamorphic grade toward the granitic batholiths indicates that higher grade metamorphism is a contact effect, but the amphibolite and upper greenschist facies assemblages are not typical contact metamorphic rocks. The mineral assemblages are similar to those produced during normal contact metamorphism, but the well developed foliation is abnormal. This foliation suggests that the metamorphism occurred in a moderate pressure regime, possibly in a stress field related to diapiric emplacement of the batholiths. The high temperature gradients at the margins of the belts (Fig. 5) and the association with batholith emplacement indicate that these assemblages are the result of mesozonal contact metamorphism rather than regional metamorphism. The discontinuous, relatively narrow, hornblende hornfels facies zone has typical contact metamorphic textures and may reflect late-stage metamorphic adjustment in a reduced stress field following batholith emplacement.

The lower greenschist and subgreenschist facies zones may reflect either burial or contact metamorphism or a combination of these effects. In the wide Abitibi belt, Jolly (1974) reported widespread subgreenschist facies, burial metamorphic assemblages. Regional greenschist facies assemblages were not observed but were predicted to occur deeper in the volcanic sequence. Contact metamorphic greenschist facies assemblages are associated with plutons.

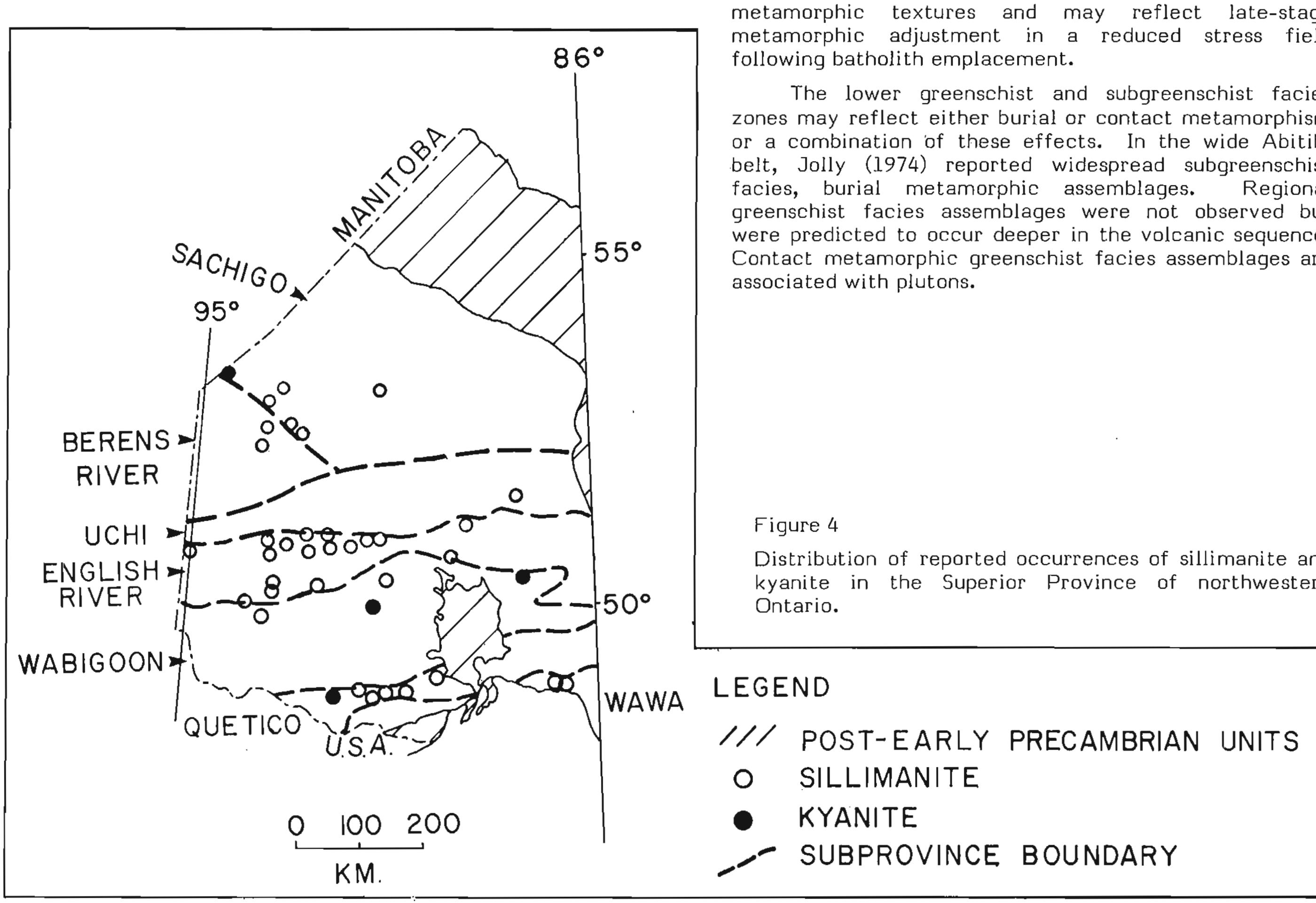

Figure 4

Distribution of reported occurrences of sillimanite and kyanite in the Superior Province of northwestern Ontario.

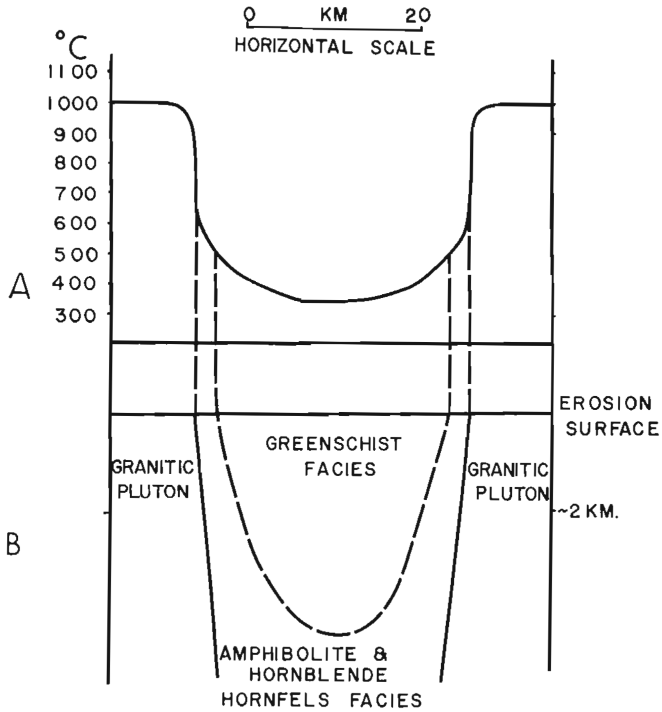

Figure 5. A. Idealized temperature distribution along a horizontal section through a greenstone belt during metamorphism related to granitic batholith emplacement. The temperature distribution will vary with depth, and this temperature distribution corresponds to the erosion surface shown in part B.

B. Cross-section through the greenstone belt showing distribution of metamorphic zones as a function of depth.

In the narrower greenstone belts of northwestern Ontario, subgreenschist facies metamorphism is apparently restricted to the Uchi Lake area (Fig. 1; Thurston and Breaks, 1978), where it may be the result of burial metamorphism. In other greenstone belts, the inner part of the greenschist facies zone in the wider belts, where horizontal temperature gradients are relatively gentle, may be the result of burial or of upgrading of burial assemblages by the batholiths. Such burial metamorphism would be the only true regional metamorphism in the greenstone belts. It should be noted that, considering the high geothermal gradient, the concentration of heat flow and thickness of the volcanic sequences, burial metamorphism would have been inevitable in the lower parts of the sequences. Furthermore, the subgreenschist facies assemblages in the Uchi Lake and Abitibi belts could have been preserved only if the greenstone belts were not deeply buried. In fact, the overall metamorphic grade of all greenstone belts and the high horizontal thermal gradients at the margins of the belts indicate relatively shallow burial.

Using the thermal gradients of Hargraves (1976) and Lambert (1976) and assuming that the boundary between greenschist and subgreenschist facies represents a temperature of about 250°C (Turner, 1968), the maximum burial depth for the wider greenstone belts was 7 km. Burial depths for most of the narrower greenstone belts, although variable, were probably comparable. In the narrower belts subgreenschist facies assemblages were probably upgraded by heat from the batholiths.

The high geothermal gradient in the early Precambrian produced subgreenschist and other burial metamorphism assemblages at shallower depths than Phanerozoic sequences of comparable lithology and metamorphic grade. Thus, burial depths in early Precambrian sequences cannot be estimated from comparison with better documented Phanerozoic sequences. Furthermore, a high geothermal gradient means that blueschist or kyanite-bearing assemblages typical of high pressure facies series would not have developed under normal conditions in early Precambrian sequences, and particularly not in greenstone belts where the geothermal gradients were highest.

The granitic batholiths are composed of numerous discrete intrusive phases with a general granodioritic to tonalitic composition (Ayres, 1974a; Ayres and Ermanovics, 1972). The specific granitic phase in contact with the greenstone belt changes from place to place along the contact with no apparent change in the metamorphic zoning. Thus the metamorphic zoning seems to be related to the emplacement of the batholith as a whole rather than to individual phases.

The emplacement temperatures of the batholiths were strongly dependent on water content and degree of crystallization. According to Wyllie et al. (1976), the water content was probably in the range 1 to 2 per cent, and the temperature of a completely molten tonalite magma could have been 1100 to 1150°C. If the magma contained some crystals, the temperatures would have been lower (Wyllie et al., 1976; Piwinskii, 1973), probably near 1000°C. This may have been the temperature of the magmatic heat source (Fig. 5), and probably was relatively constant from belt to belt. The temperature in the outermost part of the greenstone belt would have been considerably lower than the magma temperature (Jaeger, 1957), but the temperature contrast is difficult to estimate because the burial temperature of the greenstone belt at the time of batholith emplacement is unknown.

Variations in Width of Amphibolite Facies Zones

If the batholithic heat source was relatively constant, then the variations in width of the higher grade metamorphic zones must reflect other factors such as changes in lithology and stratigraphy of the greenstone belt, attitude of the batholith contact, compositional or depth differences between granitic plutons, or depth of burial.

Lithologic control can be documented in the Muskrat Dam Lake belt where several pre-metamorphic gabbro sills are present in the metavolcanic-metasedimentary sequence (Ayres, 1969b). One of the thicker sills (1200 m) in the southwestern part of the belt (locality 1, Fig. 3) has influenced the position of the amphibolite-greenschist boundary. In the western part of the sill, the distance between the sill and the granitic batholith varies from 0 to 2 km, but irregardless of this distance, the metamorphic boundary is several hundred metres inside the sill. This localized control of metamorphism may have resulted from reduced thermal conductivity in the sill because of coarser grain size, lower porosity, and lower primary fluid content. Such lithologic control probably had a relatively minor effect on metamorphism.

The effect of stratigraphy on the width of the amphibolite facies zone has been documented for the Setting Net Lake belt (Fig. 2). In the Muskrat Dam Lake belt, the west to east change in width of the amphibolite facies zone is accompanied by a lithologic change, with the eastern and western segments being separated by the Windigo River fault (Fig. 3). The eastern part, where the amphibolite facies zone is wider, is composed mainly of subaqueous mafic flows that may represent the lower portion of a typical greenstone belt

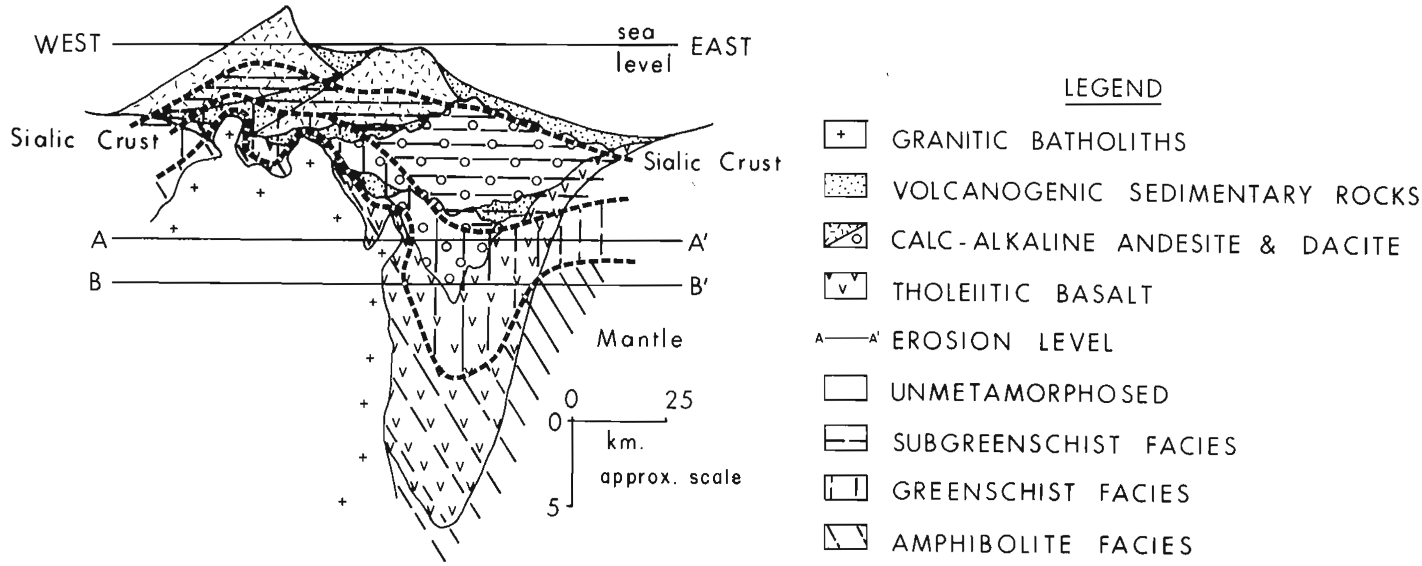

Figure 6. Idealized cross-section through a simple greenstone belt showing isostatic downwarping, deformation, and metamorphism concomitant with granitic batholith emplacement and surface volcanism. At this stage only one batholith has been emplaced. Emplacement of a second batholith on the east side of the belt would make the metamorphic zones more symmetrical. Erosion level A-A' corresponds to the Setting Net Lake belt and B-B' to the Muskrat Dam Lake belt.

volcano. The western part, where the amphibolite facies zone is narrower, is composed mainly of felsic to intermediate pyroclastic rocks and associated greywacke that may represent the upper part of the volcano. If the correlation between lithology and stratigraphic position is correct, then the lower part of the sequence, as in the Setting Net Lake belt, has a wider amphibolite facies zone.

Along the southwest edge and in the north-central part of the Muskrat Dam Lake belt (localities 3 and 4, Fig. 3), a widening of the amphibolite facies zone appears to reflect an outward bulge or re-entrant in the near vertical greenstone belt-batholith contact. In the northwest part of the Muskrat Dam Lake belt (locality 2, Fig. 3), the wider amphibolite facies zone in the fault-bounded wedge is associated with a shallower, inward-dipping contact.

In the northwest part of the Setting Net Lake area, the difference in width of the amphibolite facies zone (Fig. 2) may indicate differences in the bordering granitic plutons. The northern pluton is a typical, mesozonal batholith, whereas the southern pluton is a large zoned stock that may represent a higher emplacement level (Ayres, 1974a).

In spite of these various explanations, most width variations of metamorphic zones along the strike of the Muskrat Dam Lake belt and among other belts are not a simple function of lithologic, stratigraphic, or structural differences in the greenstone belts, nor of compositional or size variations in the batholiths. Thus the amphibolite facies zone in the Muskrat Dam Lake belt is variable in width from west to east, but is of greater width throughout than that of the Setting Net Lake belt. In part the differences are related to different stratigraphic levels in the volcanoes, but the major factor seems to be differences in depth of burial, now shown by different levels of erosion in the greenstone belts.

Increasing depths of burial in a greenstone belt may have resulted in higher initial temperatures and higher load pressures. At deeper levels, the combination of slightly higher temperatures and a slower dissipation of magmatic heat may have caused a wider amphibolite facies zone to develop. In this model the eastern part of the Muskrat Dam Lake belt represents a deeper erosional level than the western part, and the entire belt is more deeply eroded than the Setting Net Lake belt.

Regional stratigraphy supports these conclusions. Because of a higher thermal gradient and a less stable crust, early Precambrian volcanoes may have been depressed isostatically into the crust and mantle during volcanism, producing synclinal structures (Hargraves, 1976; Ayres, in prep.). The lower mafic parts tend to be depressed more than the younger, felsic to intermediate upper parts (Fig. 6). Consequently, when the granitic batholiths were emplaced, wider amphibolite facies zones developed in the lower parts of the depressed volcano (Figs. 5,6). Thus, in general, thick mafic volcanic sequences should have wider amphibolite facies zones than felsic to intermediate volcanic and sedimentary sequences. This is the relationship shown in the Muskrat Dam Lake belt. In the Setting Net Lake belt, only the upper part of the volcano is represented (Ayres, 1977). Thus the relatively narrow amphibolite facies zone in this belt also fits the depth model.

The generally high thermal conditions within greenstone belts suggests that none of the belts was deeply buried and therefore burial depths among belts cannot differ by more than a few kilometres. Also many early Precambrian volcanoes were thicker than the average burial depth. Thus the basal part of many volcanoes, at least in the central thickest part of the volcanoes, will have been depressed below the present erosion level (Ayres, in prep.).

The correlation between width of metamorphic zones, depth of burial, and stratigraphy is obviously very crude. However, it does fit available stratigraphic and metamorphic data as well as a general isostatic model for the origin of greenstone belts developed from other arguments (Ayres, in prep.). Furthermore, it shows the necessity of considering metamorphism in any model for the origin of greenstone belts. A corollary to the model is that, within subprovinces, the width of metamorphic zones may be an added tool for regional stratigraphic correlation.

Granitic Batholiths

In most areas there is no strong evidence of metamorphism of the granitic batholiths, although there is evidence of pervasive recrystallization. This can be recognized by the presence of quartz foliation, discordant cross-cutting foliation, and local recrystallization of plagioclase and mafic minerals. Hence Moxham (1965) considered the granitic rocks to be metamorphosed, whereas Ayres (1974a) concluded that they were not metamorphosed.

The most common foliation in the batholiths is due to oriented mafic minerals or aggregates of mafic minerals, and to monomineralic lenses of quartz and plagioclase. Such features are not common flow phenomena, but are probably strain effects produced by a combination of deformation and recrystallization. The late development of this foliation can be documented where several granitic phases are in contact, particularly where a younger phase forms relatively narrow dykes in an older phase. Foliation having the same orientation in both phases but discordant to the contact indicates that foliation developed after emplacement.

Development of the discordant foliation did not cause widespread recrystallization. In the batholithic rocks plagioclase has its characteristic subhedral shape with well developed oscillatory or normal zoning, although bending and other local strain effects are common. Microcline is anhedral to subhedral and generally well twinned and perthitic. The mafic minerals commonly form aggregates of biotite – epidote ± hornblende – accessory minerals and are not obviously recrystallized. Even quartz is not generally affected, although it does show variable undulatory extinction.

These textures imply magmatic crystallization, whereas the discordant foliation implies recrystallization. If the foliation developed during cooling, it would represent an adjustment to strain imposed by the country rocks, emplacement of younger phases, and possibly by post-consolidation diapiric uprise of several phases in combination. As such, the foliation is really not a metamorphic effect, but it is part of the intrusion process.

A more intense and apparently superimposed recrystallization occurs in granitic rocks of the Sachigo subprovince adjacent to the Berens River subprovince. There, a cataclastic zone separates granitic plutons of the two subprovinces. Adjacent to, and on the north side of the boundary, is an intermediate metavolcanic unit 250 to 500 m wide that is strongly recrystallized but only weakly cataclastic. Granitic plutons of the Sachigo subprovince north of the metavolcanic unit are not obviously cataclastic and superficially resemble other granitic plutons of the subprovince. Microscopically however, they show evidence of recrystallization in a zone 5 km wide adjacent to the boundary. Biotite is recrystallized to fine grained aggregates of biotite and epidote, and the feldspars and quartz are more recrystallized than in other granitic plutons of the subprovince. This recrystallization appears to be the result of movement along the boundary fault.

Metasedimentary xenoliths in the Sachigo subprovince batholiths show a progressive increase in metamorphic grade towards the Berens River subprovince boundary. Away from the boundary the characteristic mineral assemblage in xenoliths of appropriate composition is cordierite – andalusite, but in the 5 km wide boundary zone sillimanite is a stable phase.

Cataclasis is concentrated in the granitic rocks of the Berens River subprovince and ranges from ultramylonite at the boundary to weak cataclasis 2 km south of the boundary. The concentration of cataclasis on the south side of the boundary probably indicates a deeper crustal level for the Berens River subprovince. Relative vertical movement of the two subprovinces, based on the nature of the plutons and metamorphic grade, is south side up.

Small synvolcanic granitic plutons, which occur in some greenstone belts, have been metamorphosed to the same grade as adjacent country rocks. In the amphibolite facies zone, recrystallization is extensive and the metamorphic effects can be readily recognized, particularly at the margins where strain effects and fluid ingress are most pronounced. In the greenschist facies zone, metamorphic effects are readily recognizable in fine grained plutons but not in medium grained plutons. In the latter the massive nature and medium grain size inhibit recrystallization; metamorphic minerals are patchily developed; and those that do form, such as chlorite, albite, epidote, carbonate, and muscovite, cannot always be distinguished from deuteric alteration assemblages. Because of the difficulty in recognizing metamorphic effects in medium grained plutons some metamorphosed synvolcanic plutons probably have not been recognized.

Sialic Crust

Remnants of pre-volcanism sialic crust have been recognized in three subprovinces. The distribution of the remnants suggests that the crust was once widespread (Ayres, 1974a, 1974b; Harris and Goodwin, 1976), but it is now dismembered and engulfed by younger granitic plutons. In the Setting Net Lake area of the Sachigo subprovince, a zircon U-Pb age of 2950 Ma has been obtained from a metamorphosed trondhjemite unit (Krogh and Davis, 1971; Ayres, 1974a; Hillary, 1976). In the English River subprovince, a zircon age of 3040 Ma has been obtained from metamorphosed tonalite (Krogh et al., 1976b), and similar ages have been obtained from comparable units in the Berens River subprovince of Manitoba and Ontario (Krogh et al., 1976a). Similar, but as yet undated, rock units are present elsewhere in northwestern Ontario. These ages are 200 to 300 Ma older than the volcanic sequences (Krogh and Davis, 1971).

The crustal remnants have been recrystallized (Harris and Goodwin, 1976; Hillary, 1976) to amphibolite and locally granulite facies, although diagnostic mineral assemblages are rare. Features of these remnants that support an original plutonic origin include: 1) igneous composition, 2) uniformity of composition over relatively large areas, 3) relict igneous textures in the least recrystallized portions, and 4) the presence of euhedral to subhedral zircon characteristic of magmatic rocks (Harris and Goodwin, 1976).

BERENS RIVER SUBPROVINCE

The Berens River subprovince, less well known than the others, is bounded by major cataclastic zones. Granitoid rocks comprise about 90 per cent of the subprovince, and greenstone belts, as much as 8 km wide, the remainder. These greenstone belts, although less abundant, are generally similar to those of the bordering Sachigo and Uchi subprovinces except that metamorphic grade is higher. All greenstone units are within the amphibolite facies, and sillimanite is a common mineral in metasedimentary rocks (Fig. 4).

The origin of the granitoid rocks is controversial. In Manitoba, a gneissic granodiorite unit forms 46 per cent of the subprovince and may represent migmatized and recrystallized paragneiss (Ermanovics, 1971; Ermanovics and Davison, 1976). This gneissic granodiorite unit decreases in abundance eastward (Ermanovics, 1971) and is rare in northwestern Ontario, where the granitoid rocks are largely composite batholiths that range in composition from quartz diorite to quartz monzonite. These rocks are medium to coarse grained and foliated to gneissic. As in batholiths of the greenstone-granodiorite subprovinces, much of the foliation is secondary, and strong recrystallization is restricted to cataclastic zones.

The higher metamorphic grade and relative paucity of greenstone belts suggest that the Berens River subprovince represents a deeper and more plutonic crustal level than the bordering Sachigo and Uchi subprovinces. The Berens River subprovince, presumably uplifted several kilometres along boundary faults, is not just a more deeply eroded equivalent of the adjacent subprovinces. Along part of its north margin granitic phases parallel the boundary fault (Ayres, 1970), suggesting that the fault represents a pre-existing boundary. Granitic phases in the Sachigo subprovince, on the other hand, seem to be truncated by the fault. The nature and origin of the primary boundary is unknown but the relationship between the granitic phases on either side of the boundary suggests that plutonic activity in the Berens River subprovince may be younger than that in the Sachigo subprovince.

ENGLISH RIVER AND QUETICO SUBPROVINCES

Metamorphism in the English River and Quetico subprovinces has been described by Bau (1976), Beakhouse (1977), Breaks and Bond (1977), Harris and Goodwin (1976), Irvine (1963), Kehlenbeck (1976), Pirie and Mackasey (1978), and Thurston and Breaks (1978). These subprovinces differ from the other subprovinces in both lithology and metamorphism. Metasedimentary rocks are the main supracrustal components, whereas metavolcanic rocks are rare. Granitic plutons form less than 50 per cent except in the southern Winnipeg River portion of the English River subprovince where they predominate. The plutons range from strongly recrystallized orthogneiss that may be older sialic crust (Harris and Goodwin, 1976) to unrecrystallized postkinematic granitic rocks.

In the metasedimentary and orthogneiss components, metamorphic grade is predominantly middle to upper amphibolite facies although greenschist, lower amphibolite, and granulite facies assemblages are locally present (Breaks and Bond, 1977; Harris and Goodwin, 1976). Metamorphic gradients are relatively gentle without the rapid metamorphic transitions that characterize the greenstone-granodiorite subprovinces.

Greenschist facies assemblages are most common near the margins of the subprovinces in areas where metasedimentary rocks are in contact with greenschist facies greenstone belts in the adjacent subprovinces (Bau, 1976; MacDonald, 1944). In these areas there is a progressive increase in metamorphic grade across the boundary towards the higher grade assemblages in the metasediments. In many places, however, where a fault marks the contact between the metasedimentary deposits and the adjacent greenstone belts, there is an abrupt increase in metamorphic grade across the fault; in this case also the higher grade assemblages are in the metasediments. Granulite facies assemblages are only locally developed and appear to be restricted to the English River subprovince.

The metasediments are characterized by anatectic melting and the development of migmatite (Breaks and Bond, 1977). Sillimanite is common in the paleosome. Sparse clinopyroxene – garnet assemblages indicate local attainment of granulite facies conditions (Harris and Goodwin, 1976).

The orthogneiss plutons are strongly recrystallized and deformed and have a well developed gneissosity produced by a combination of metamorphism, anatexis, and magmatic injection (Breaks and Bond, 1977). Orthopyroxene-bearing assemblages indicative of granulite facies are locally present (Harris and Goodwin, 1976).

The other plutonic rocks are much less recrystallized than the orthogneiss. The older sodic plutons have well developed metamorphic foliation and are locally migmatitic (Breaks and Bond, 1977); they were apparently emplaced during the major deformational and metamorphic event. The younger potassic plutons are only locally recrystallized and postdate much of the regional metamorphism.

The relatively high abundance of granitic plutons in the greenstone-granodiorite subprovinces compared with the metasedimentary portions of the English River and Quetico subprovinces suggests that the granitic plutons are both spatially and genetically related to the volcanism. The relatively high abundance of plutons associated with orthogneiss in the southern portion of the English River subprovince is anomalous. Possibly some of the plutonism in this area predates the sedimentation and may be related to the initial deformation and metamorphism of the orthogneiss.

As previously discussed, the English River and Quetico subprovinces appear to represent deformed linear turbidite basins, in part overlying sialic basement, that formed adjacent to volcanic island chains. If these subprovinces and the adjacent greenstone-granodiorite subprovinces represent the same crustal level, then the higher metamorphic grade in the English River and Quetico subprovinces is anomalous. Volcanism and associated plutonism would have concentrated heat flow in the adjacent greenstone-granodiorite subprovinces which thus should have the higher metamorphic grade. The metamorphic anomaly must indicate uplift of the English River and Quetico subprovinces relative to adjacent subprovinces, either along faults at, or near the primary lithologic boundaries, or by upwarping. The occurrence of boundary faults and abrupt metamorphic transitions support uplift along faults, whereas the local preservation of metamorphic gradients indicates that some upwarping did occur. Possibly uplift was initiated by upwarping during or shortly following metamorphism and was completed by faulting. Whatever the cause of the uplift, the English River and Quetico subprovinces represent deeper crustal levels than the adjacent metavolcanic-granodiorite subprovinces. The absence of strong metamorphic gradients is characteristic both of lower crustal levels and more uniform heat flow.

CONCLUSIONS

In the Superior Province of northwestern Ontario, seven subprovinces are defined on lithologic differences, and each subprovince appears to represent a primary stratigraphic sequence. The distribution of metamorphic facies and mineral assemblages differs among the subprovinces, and the metamorphic differences are critical factors in deciphering the evolution of the subprovinces and of the Superior Province. Although metamorphic reactions and grades are poorly documented, some tentative conclusions can be made, particularly with reference to the greenstone belts.

In the early Precambrian, geothermal gradients were at least twice as high as during the late Phanerozoic, and the interpretation of metamorphism is more difficult than in comparable Phanerozoic sequences. Most metamorphic reactions are mainly a function of temperature, and in the early Precambrian such reactions would have occurred at higher crustal levels than in comparable Phanerozoic sequences. Consequently, the distribution of metamorphic mineral assemblages cannot be used as an estimate of burial depth, if Phanerozoic sequences are used as a guide.

The scarcity of kyanite probably reflects the high geothermal gradients that prevailed during the early Precambrian, particularly in greenstone belt terranes. Because of the higher geothermal gradients, high-pressure facies series assemblages comparable to those in Phanerozoic sequences are not likely to have developed.

The Wawa, Wabigoon, Uchi, and Sachigo subprovinces appear to represent metamorphosed, deformed linear volcanic island chains. The concentration of volcanism and plutonism should have increased heat flow in these subprovinces and resulted in a high and uniform metamorphic grade, but this did not occur. Instead the wider greenstone belts have a greenschist facies core surrounded by relatively narrow amphibolite and hornblende hornfels facies zones. The highest metamorphic grades are adjacent to the granitic batholiths that were the heat source for much of the metamorphism. The metamorphism is thus contact rather than regional metamorphism.

The relatively low metamorphic grade in the central parts of the wider greenstone belts may be the result of regional burial metamorphism, although definite burial metamorphic assemblages have been recognized only locally. The low metamorphic grade and the high, horizontal, thermal gradients at the margins of the belts indicate relatively shallow burial. The high geothermal gradients imply that the burial depth of most greenstone belts was less than 10 km.

The widths of the higher grade metamorphic zones vary among belts. An increase in the width of the zones apparently reflects greater burial depths. Greenstone belts with wide amphibolite facies metamorphic zones probably represent a lower stratigraphic sequence than belts with narrow zones. The width of amphibolite facies zones may be thus a crude measure of both burial depth and stratigraphic position.

The English River and Quetico subprovinces represent deformed linear turbidite basins, in part overlying sialic basement, that developed between the volcanic island chains. These subprovinces are characterized by higher metamorphic grades, gentler metamorphic gradients, and fewer granitic plutons than the greenstone-granodiorite subprovinces. They represent a deeper crustal level than the greenstone-granodiorite subprovinces, and were apparently uplifted by a combination of warping and faulting.

The Berens River subprovince also represents a deeper crustal level. It is composed mainly of granitic plutons with a few narrow greenstone belts that record higher metamorphic grades than greenstone belts in the adjacent Uchi and Sachigo subprovinces. The Berens River subprovince is not just a deeper equivalent of the Uchi and Sachigo subprovinces. The plutonism in the Berens River subprovince appears to be younger than that in the adjacent parts of the Sachigo subprovince.

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PRELIMINARY EXAMINATION OF REGIONAL METAMORPHISM IN PARTS OF QUETICO METASEDIMENTARY BELT, SUPERIOR PROVINCE, ONTARIO

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Pirie, J. and Mackasey, W.O., Preliminary examination of regional metamorphism in parts of Quetico metasedimentary belt, Superior Province, Ontario; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 37-48, 1978.

Abstract

The Quetico Belt comprises a major regional sedimentary trough over 800 km long, set between the Wabigoon and Shebandowan-Wawa metavolcanic and batholithic terranes of the Superior Province. The sedimentary rocks are a monotonous sequence of wacke-mudstone turbidites which have been deformed, progressively metamorphosed and migmatized from the margins to the centre of the belt.

In the lowest metamorphic grades sericite – chlorite is the stable pelitic assemblage. Inward from the belt margins, biotite stabilizes and both chlorite and sericite decrease in amount and disappear before almandine garnet is observed. Towards the centre of the belt the relative positions of pelitic assemblages suggest a prograde sequence of reactions in the following order:

a) *muscovite + staurolite + quartz = biotite + garnet + andalusite + water*

b) *andalusite = sillimanite*

c) *staurolite + quartz = cordierite + garnet + sillimanite + water*

d) *K-feldspar + Na-feldspar + quartz = melt*

e) *muscovite + Na-feldspar + quartz = K-feldspar + sillimanite + melt*

f) *biotite + sillimanite + Na-feldspar + quartz = K-feldspar + garnet + cordierite + melt*

Using the P-T phase diagram of Carmichael, this would indicate a path of regional metamorphism extending from extremely low greenschist facies conditions along the belt margins through points at 550°C at 3 kb and 700°C at 4 kb to a maximum of over 720°C at 4.2 kb.

Résumé

La zone de Quetico comprend un fossé sédimentaire régional de grande dimension, dont la longueur dépasse 800 km, et qui est situé entre les terrains métavolcaniques et batholithiques de Wabigoon et Shebandowan-Wawa, dans la province du lac Supérieur. Les roches sédimentaires sont une succession monotone de turbidites à wacke et mudstone qui ont été déformées, progressivement métamorphisées et migmatisées des marges au centre de la zone.

Aux degrés de métamorphisme les plus bas, l'assemblage à séricite et chlorite est l'assemblage pélitique le plus stable. Des marges de la zone étudiée vers l'intérieur, la biotite se stabilise et la chlorite et la séricite se raréfient et disparaissent avant le grenat almandin. Vers le centre de la zone, les positions relatives des assemblages pélitiques semblent indiquer une succession prograde de réactions, dans l'ordre suivant:

a) *muscovite + staurolite + quartz = biotite + grenat + andalusite + eau*

b) *andalousite = sillimanite*

c) *staurolite + quartz = cordiérite + grenat + sillimanite + eau*

d) *feldspath potassique + feldspath sodique + quartz = phase liquide*

e) *muscovite + feldspath sodique + quartz = feldspath potassique + sillimanite + phase liquide*

f) *biotite + sillimanite + feldspath sodique + quartz = feldspath potassique + grenat + cordiérite + phase liquide*

En utilisant le diagramme de phases de Carmichael avec P et T (Carmichael, 1978) on découvre une zone de métamorphisme régional, dont le domaine s'étend d'un faciès des schistes verts de degré métamorphique extrêmement faible le long des marges de la zone, en passant par des points qui ont été soumis à des températures de 550°C à 3 kb et 700°C à 4 kb, jusqu'à des points qui ont traversé un maximum dépassant 720°C à 7.2 kb.

INTRODUCTION

The metasedimentary Quetico Belt comprises a major regional Early Precambrian trough between the Wabigoon and Shebandowan-Wawa metavolcanic and batholithic terranes of the Superior Province (Fig. 1). It extends 800 km eastwards from Minnesota through the Rainy Lake area, under the Late Precambrian rocks of the Nipigon basin, to the Kapuskasing Structure south of Hudson Bay (McGlynn 1970, p. 66).

The nature and location of the boundary between the Quetico and the Wabigoon and Shebandowan belts is controversial, but most of the evidence indicates that the deposition of the sediments in the Quetico Belt was in part at least contemporaneous with the volcanism in the adjacent terranes (Goodwin et al., 1972; Mackasey et al., 1974). In places the marginal metavolcanics and nearby metasediments face the central part of the Quetico Belt. However, isoclinal folding in the metasediments complicates the stratigraphic relationships.

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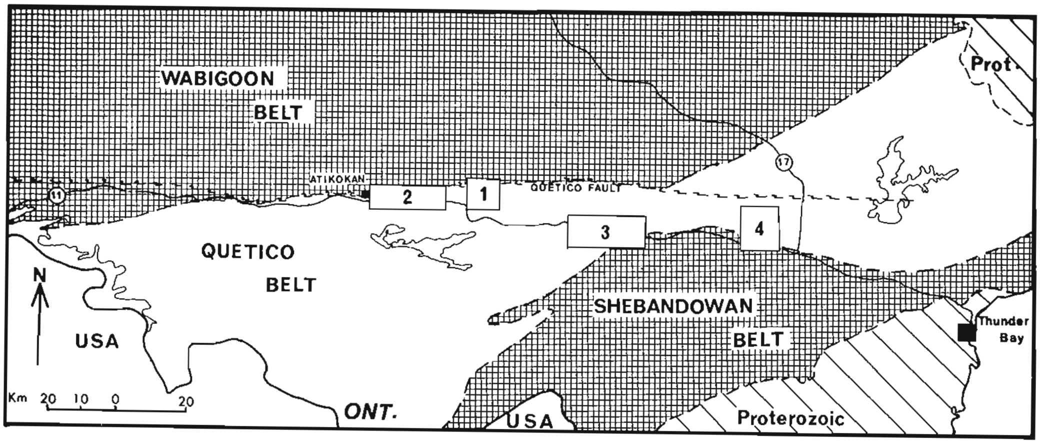

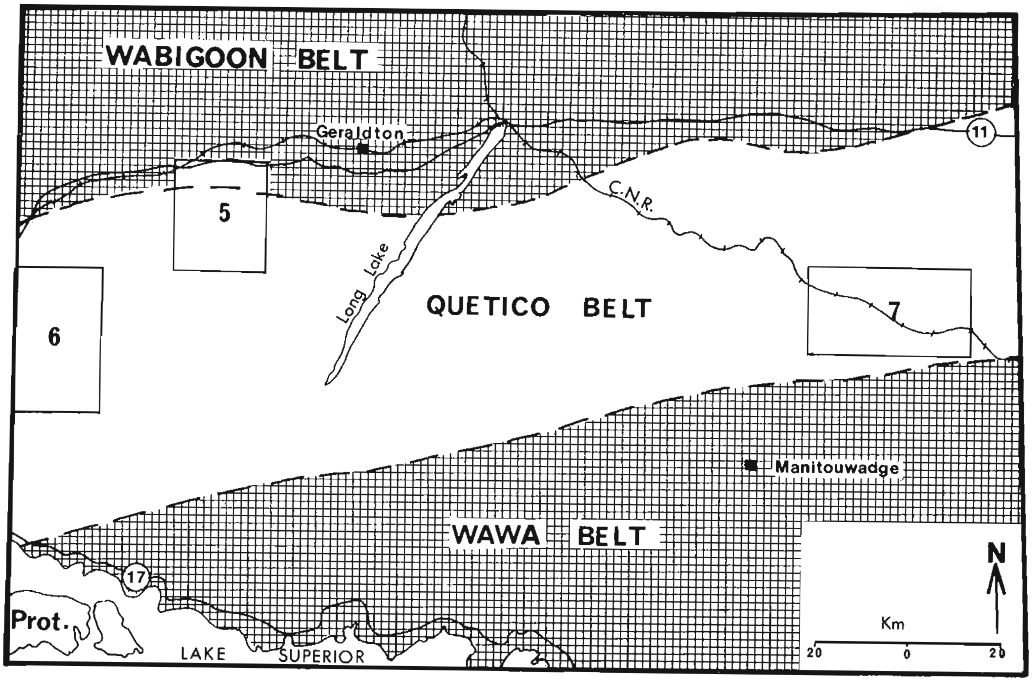

Figure 1.

1. Crooked Pine Lake Area
2. Sapawe-Atikokan Area
3. Planet-Huronian Lake Area
4. Shebandowan-Drift Lake Area
5. Jellicoe Area
6. Georgia Lake Area
7. Flanders Lake Area

Location of areas studied within Quetico metasedimentary belt, Superior Province.

The Quetico Belt comprises a rather monotonous sequence of thin to medium bedded turbidites which have been deformed, progressively metamorphosed, and migmatized from the margins inwards so that the original textures and structures of the turbidites have been erased except in marginal zones a few kilometres wide.

This paper discusses the general nature of regional metamorphism along a number of sections extending from the margins into the central higher grade migmatite zones of the Quetico Belt (Fig. 1). Some detailed geologic and metamorphic relationships, documented in the Crooked Pine Lake area east of Atikokan, are compared with reconnaissance information from other areas to provide an overview of this belt. Pelitic assemblages in the Quetico Belt are listed in Table 1.

Table 1

Pelitic assemblages observed in Quetico metasediments. All assemblages contain plagioclase and quartz

1.	chlorite – sericite – carbonate
2.	chlorite – sericite
3.	biotite – sericite – chlorite
4.	biotite – chlorite
5.	biotite
6.	biotite – garnet
7.	biotite – garnet – andalusite – chlorite
8.	biotite – garnet – andalusite
9.	biotite – garnet – andalusite – staurolite
10.	biotite – garnet – staurolite
11.	biotite – garnet – cordierite – staurolite – sillimanite
12.	biotite – garnet – sillimanite
13.	biotite – garnet – cordierite – sillimanite
14.	biotite – sillimanite
15.	andalusite – sillimanite (fibrolite)
16.	biotite – garnet – cordierite – staurolite

CROOKED PINE LAKE AREA

Metasediments

In the Crooked Pine Lake area (Pirie, 1977) the Quetico Fault, probably a right-lateral wrench fault, forms the boundary between the Wabigoon and Quetico belts. South of the fault is a steeply dipping to vertical east trending sequence of wacke-mudstone turbidite units. Where primary sedimentary structures are preserved, stratigraphic tops consistently face north so that the youngest part of the sequence is adjacent to the Quetico Fault (Fig. 2). Clastic features and sedimentary structures become blurred southwards and approximately 1.5 km south of the fault disappear due to recrystallization and deformation.

The turbidite beds are sharp-based and generally between 5 and 40 cm thick. The lower part of a bed consists of light to dark grey, graded or massive, weakly foliated wacke merging into a darker, fissile siltstone or less commonly, shale top. In many localities, it is difficult to determine whether the planar structures whithin the beds are original sedimentary laminations or foliation planes related to later deformation, although locally the later foliation is at a slightly oblique angle to the lamination and bedding. In some places beds contain little or no mudstone, whereas in others thinly laminated siltstone units commonly grade into shale. Minor, thin arkosic layers occur sporadically throughout the sequence and in a few places at the west end of Crooked Pine Lake there are pebble to granule conglomerate beds up to 30 cm thick. These conglomerates contain mainly elongate, fine grained, intermediate to felsic volcanic clasts, up to 4 cm long set in a wacke matrix.

The beds, composed of lower massive or graded wacke and upper laminated mudstone, are typical of turbidites and represent the A, B, and D parts of the Bouma sequence (Bouma, 1962). Locally, current-bedding, dewatering structures, and ripple marks are present. Northeast of Kawene Lake, in turbidite layers less than 10 cm thick, stratiform zones of intraformational breccia and slump structures including minor, open, soft sediment folds and miniature thrust faults, are present.

In the zone of lowest metamorphic grade south of the Quetico Fault, the wacke components of the turbidites contain variable proportions of quartz, plagioclase, lithic clasts, and locally carbonate ranging from 0.5 to 1.5 mm in diameter in a 30 to 60 per cent silty matrix. These turbidites fall within the broad range of lithic arkosic wacke and lithic subarkosic wacke classification (Young, 1967). The mineral and lithic clasts are rounded to subangular. In both clast types quartz is strained and plagioclase albite-twinned and lightly dusted with sericite, but the clasts have undergone little recrystallization compared with the sericite-chlorite matrix.

The mudstone layers are well laminated and foliated, and contain very fine grained quartz, plagioclase, sericite, chlorite, and locally epidote and carbonate. The laminations, typically less than 5 mm thick, consist of alternating micaceous shale and quartzofeldspathic siltstone.

Regional Metamorphism

South of Quetico Fault and Crooked Pine Lake, the grade of metamorphism in the metasediments increases over a distance of 5.5 km from low greenschist facies, just south of the Quetico Fault, to upper amphibolite facies with concomitant anatexis at Eva Lake. Metamorphic isograds (Fig. 2) trend easterly parallel to the main layering and foliation in the turbidite metasediments. The combination of variable outcrop pattern, variation in bulk composition of pelitic layers, intrusion of large pegmatite bodies, and difficulty in positively identifying phases such as cordierite in the field, all mitigate against complete unravelling of the metamorphic relationships.

Typical pelitic assemblages from the Crooked Pine Lake area (Table 1, Assemblages 1-14) are given in approximate order of increasing grade from north to south. The locations of some of these assemblages are shown in Figure 2. Because of the absence of muscovite in the higher grade assemblages due presumably to the low initial potassium content, and the exhaustion of sericite in the lower grade reaction, few of the typical pelite mineral reaction isograds can be illustrated. The first and last occurrences from north to south of particular minerals have been noted, in addition to reactions which fit the assemblage data. In the more pelitic tops of turbidite units around the south shore of Crooked Pine Lake, chlorite, sericite, and carbonate accompany plagioclase and quartz. A short distance to the south biotite appears, chlorite and sericite decrease in amount, and about 500 m south finally disappear (Assemblages 1 to 5, Table 1). These relationships suggest a reaction involving sericite and chlorite to form biotite, possibly as proposed by Winkler (1976, p. 214):

phengite + chlorite = biotite + Al richer chlorite + quartz

In the vicinity of Crooked Pine Lake, albite-twinned plagioclase clasts are lightly dusted with sericite and within one sample may range in composition from An_{15} to An_{30}, reflecting variety in source material.

Garnet first appears in the rocks south of an east-trending line through the south end of Tower Lake (Fig. 2) and about 90 m to the south, staurolite and andalusite appear with garnet and biotite. The apparent geographic coincidence in the appearance of these minerals may be in part due to variations in bulk rock compositions. The suppression of the low temperature parts of the garnet and staurolite stability fields, for example, may be related to the lack of available muscovite to combine with chlorite and quartz to give the typical almandine garnet- and staurolite-forming reactions.

The assemblages containing biotite – garnet – andalusite (Assemblages 7, 8, 9, Table 1, and Fig. 2) are constrained by the reaction curve

muscovite + staurolite + quartz = biotite + garnet + Al_2SiO_5 + water (R.1)

to the high temperature side and to the area below the andalusite-sillimanite boundary (Fig. 3).

Assemblages containing staurolite (Assemblages 9, 10, 11, Table 1 and Fig. 2) occur over a distance of about 1600 m and are constrained to the high temperature side of a reaction (Fig. 3) such as

muscovite + chlorite + almandine = quartz + biotite + staurolite + water (R.2)

and the low temperature side of the reaction (Fig. 3)

staurolite + quartz = cordierite + garnet + Al_2SiO_5 + water (R.3)

Assemblage 11 (Table 1 and Fig. 2) observed just east of Kawene station would lie on this reaction isograd and in fact thin sections show the presence of only small local staurolite "droplets" generally set in plagioclase, with cordierite, garnet, and fine sillimanite nearby. The curves for reactions (1) and (3) are close together in Figure 3, but they are arranged compatible with the sequence of mineral assemblages observed in the field. As is evident from the assemblages in Table 1, the stable aluminum silicate polymorph also changes somewhere between reactions (1) and (3) (geographically on the map along a line trending roughly eastwards through the southern parts of Kawene and Heward lakes in Fig. 2). The relationships outlined on Figure 3 suggest that Assemblage 11 (Table 1) was formed at about 550°C and a pressure of about 3 kb[1].

[1] 1 kb = 1×10^5 kPa

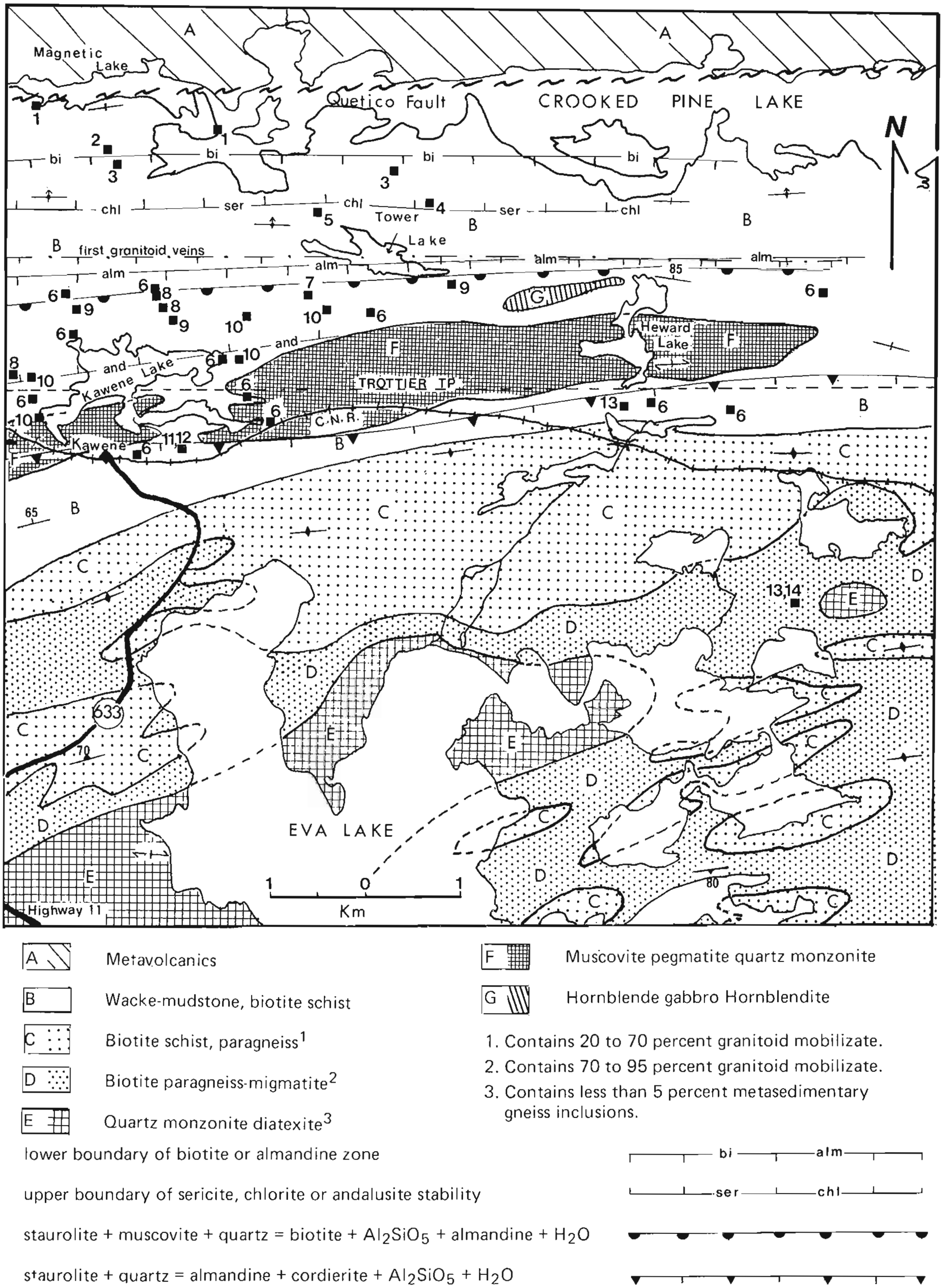

Figure 2. Geology of Crooked Pine Lake area with approximate metamorphic isograds and zones. Locations of pelitic mineral assemblages are numbered as in Table 1.

The assemblage biotite + garnet + cordierite + sillimanite (Assemblage 13, Table 1, Fig. 2) was noted just south of Heward Lake; this reaction isograd appears to extend due east from Kawene station through the southern tip of Heward Lake. Few localities with pelitic assemblages were found farther south partly because of poorer outcrop and the sharp increase in quartz monzonite and pegmatite which becomes the predominant component of the paragneiss-migmatite some 2000 m south of Kawene and Heward lakes. Assemblage 14 (Table 1, Fig. 2) comprises aggregates of coarse sillimanite and fibrolite accompanying biotite. Nearby, Assemblage 13, biotite – garnet – cordierite – sillimanite, was observed. These assemblages are constrained to the low temperature side of the reaction

biotite + sillimanite + Na feldspar + quartz =
K-feldspar + garnet + cordierite + melt (R.4)

and the high temperature side of the minimum melting curve for anatexis (Fig. 3); quartz monzonite is the predominant component in the surrounding migmatites.

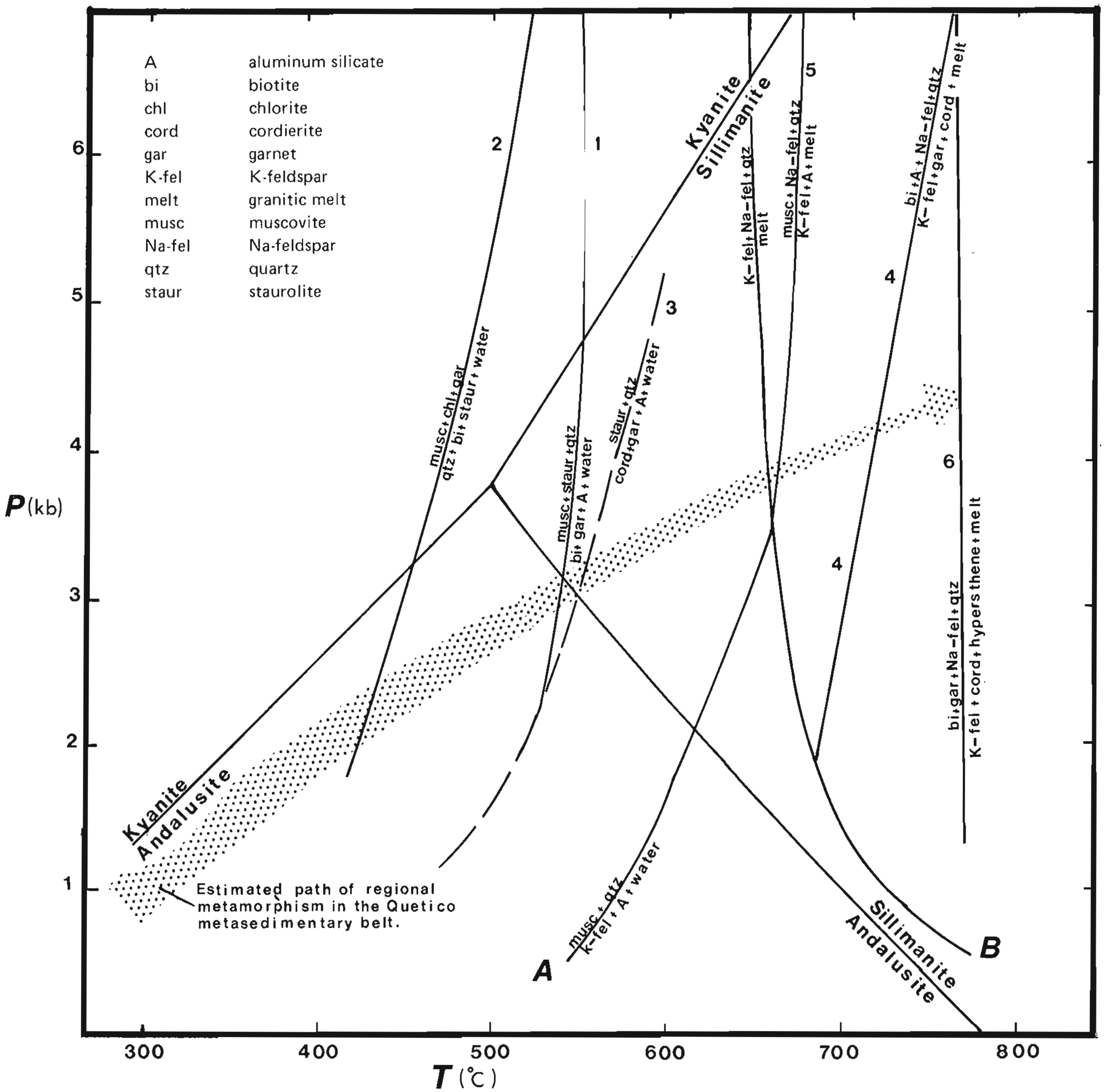

Figure 3. P-T phase diagram for part of the "ideal" pelitic system SiO_2-Al_2O_3-FeO-MgO-Na_2O-K_2O-H_2O and various subsystems relevant to the metamorphism of pelitic schist and gneiss containing quartz and muscovite or K-feldspar (Dashed reaction curve is muscovite deficient). Taken with permission from the more complete P-T diagram of D.M. Carmichael (pers. comm.) constructed using Holdaway (1971) for aluminum silicate stability fields. Chatterjee & Johannes (1974) for curve (A), and Kerrick (1972) for curve (B). Relationships between these and the other curves are based on field observations, partition coefficient data, and theoretical chemographic considerations.

The granitoid rocks also show progressive changes from north to south; quartz monzonite and pegmatite veins and stringers first appear north of the garnet isograd (Fig. 2). The proportion of granitoid material in general increases steadily southwards (with the anomalous exception of the large muscovite pegmatite and quartz monzonite sheet extending eastwards through Kawene and Heward lakes), forming larger sheets and bodies as well as more numerous small scale concordant layers and veins within the biotite paragneiss until the quartz monzonite material becomes the predominant phase, and contains rafts and blocks of the layered biotite paragneiss. Finally, large elongate bodies of homogeneous slightly foliated quartz monzonite diatexite more than 1 km wide occur within zones of biotite gneiss-migmatite around Eva Lake. Estimates of granitoid content in the metasediments were made at outcrop scale during mapping using arbitrary cut-off percentages of 20, 70, and 95, to indicate increasing degrees of migmatization and anatexis (Fig. 2). A plot of modal mineralogy for samples of granitoid mobilizate in the paragneiss, gneiss-migmatite, and homogeneous diatexite (Fig. 4) shows a close relationship between the modes of the rocks and the minimum melting compositions in the granite system outlined by Tuttle and Bowen (1958).

Carbonate-bearing layers are not common in the turbidites and are difficult to distinguish from the typical semipelites exposed around Crooked Pine Lake. However, as the metamorphic rank increases pelitic layers are characterized by the development of biotite, and carbonate layers by prophyroblasts of actinolite amphibole. In the migmatite zone to the south, equivalent Ca-rich rocks are hornblende biotite schists. The only occurrences of thin marble units are at the south end of Elbow Lake some 6 km east of Eva Lake. There calc-silicate rocks comprising actinolite, diopside, grossularite, wollastonite, and calcite are associated with pelites metamorphosed to upper amphibolite facies and with migmatites containing a high percentage of anatectically derived quartz monzonite.

SAPAWE-ATIKOKAN AREA

The eastern half of this area (Fig. 5) was mapped by McIlwaine and Hillary (1974) and the area around Atikokan was included in the Steep Rock Lake area mapped by Shlanka (1972). Examination of roadside outcrops shows that the regional metamorphic pattern is quite similar to that of the Crooked Pine Lake area about 7 km to the east. A sequence of assemblages similar to that in the Crooked Pine Lake area extends southward from the Quetico Fault. In general, however, the lower grade phases in the Sapawe-Atikokan area are stable across greater widths. The assemblage biotite – garnet – andalusite noted at two localities on Highway 11 constrains the isograd for reaction (1) to a location somewhat north of these localities (Fig. 5). The spatial relationships between the biotite "in", sericite and chlorite "out" lines vary across the area and are no doubt controlled by bulk rock composition.

PLANET-HURONIAN LAKES AREA

This area (Fig. 6), mapped by Giblin (1964) and Harris (1970), is underlain by metasedimentary and migmatitic rocks similar to those in the Crooked Pine Lake area. The east-northeast trending contact between mafic metavolcanics of the Shebandowan Belt to the south and east and the wacke-mudstone turbidites of the Quetico Belt is exposed on Highway 11. Pillow top and grain gradation determinations at this boundary indicate the sequence faces northerly although about 1 km west the rocks face south, suggesting a synclinal structure within the metasediments near the metavolcanic contact, with a complementary anticline towards the north. Minor mafic metavolcanic units are interbedded with the metasediments for a short distance into the Quetico Belt, providing evidence that the boundary between Quetico and Shebandowan belts in this region is more a facies change than a major structural discontinuity.

Metamorphic zones in the Planet-Huronian Lake area are assumed to be approximately parallel to the main lithologic units and bedding (Fig. 6). The metamorphic grade in the metasediments at the southern boundary of the Quetico Belt is higher than the grade in the metasediments adjacent to the Quetico Fault on the northern boundary; biotite in coarse randomly oriented porphyroblasts and smaller aligned flakes is common in the metasediments adjacent to the pillowed mafic metavolcanics of the Shebandowan Belt. The possibility that the metavolcanics have also undergone contact metamorphism following intrusion of a granitic pluton about 1 km to the south cannot be dismissed. The upper boundaries of sericite and chlorite zones approximately coincide in the metasediments about 500 m from the main

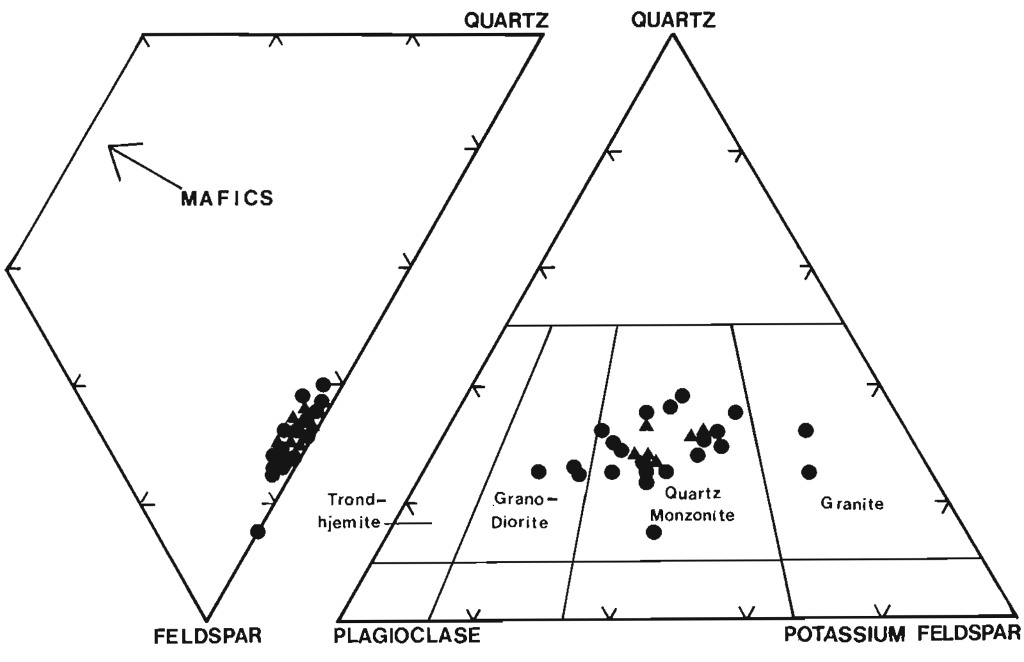

Figure 4.

Triangular plots of modal mineralogy of granitoid mobilizate in gneiss, gneiss-migmatite (circles), and quartz monzonite diatexite (triangles) from Crooked Pine Lake area (Modal counts on stained slabs using grid of 10 lines/inch and a count of 500 points).

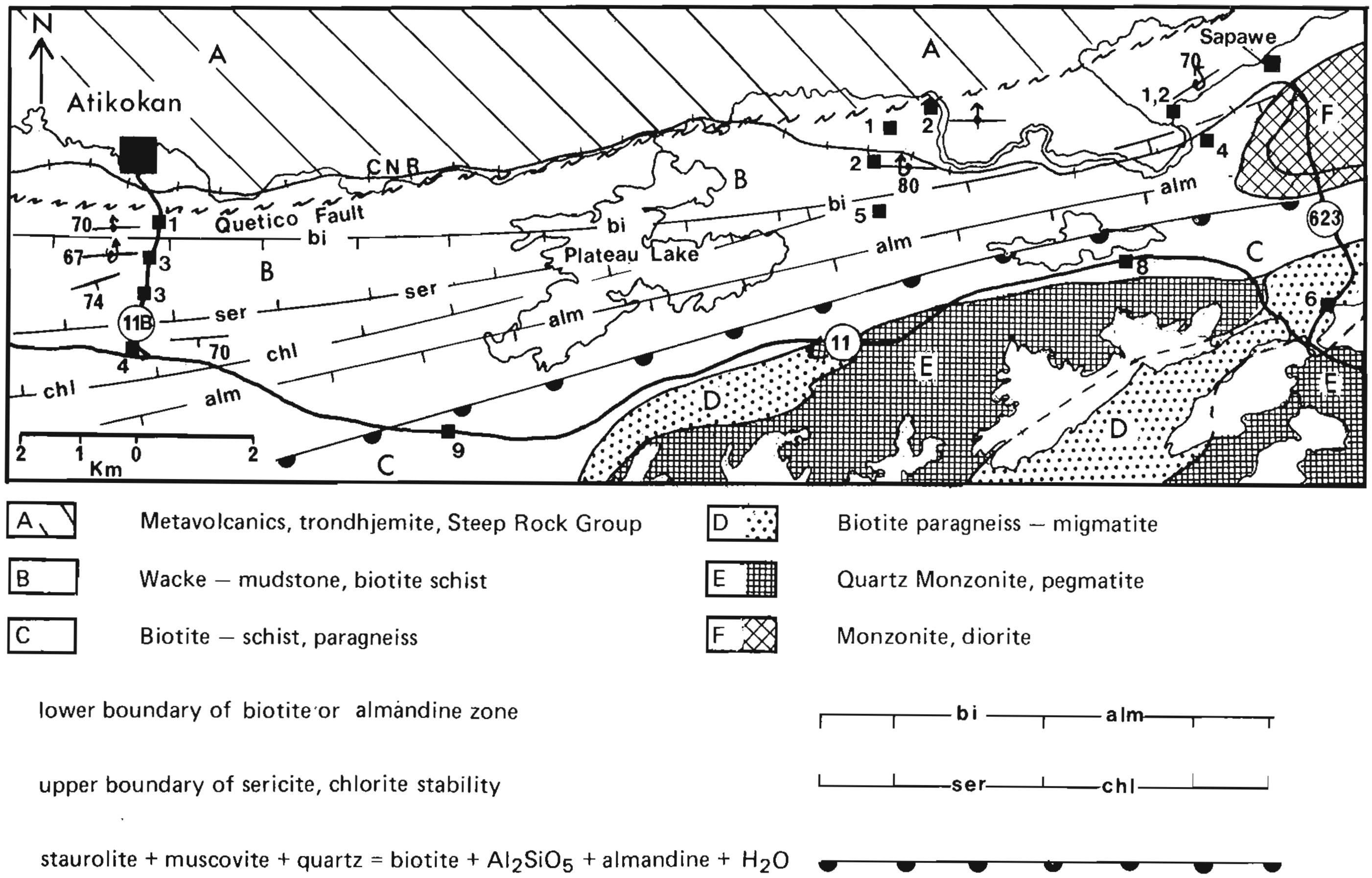

Figure 5. Geology of Sapawe-Atikokan area with approximate metamorphic zones and isograds. Location of pelitic mineral assemblages are numbered as in Table 1.

metavolcanic contact and almandine garnet appears 500 m farther northwest (Fig. 6). Some 8 km by road west of the main metasediment-metavolcanic contact, pelitic layers contain the assemblages biotite – garnet – staurolite – andalusite – fibrolite (cf. Assemblage 9, Table 1) and biotite – garnet – staurolite – cordierite – andalusite – fibrolite (cf. Assemblage 11, Table 1). The coexistence of the two aluminum silicate polymorphs is of particular interest. Fibrolite commonly grows on biotite or is closely associated with minor late muscovite, but in the Assemblage 9 equivalent, the fibrolite was observed within the andalusite. Holdaway (1971) considered fibrolite always to be a metastable phase forming from mica with minor overstepping into temperatures slightly higher than those of the predicted andalusite-sillimanite transition. On the basis of energy requirements Holdaway ruled out the formation of fibrolite directly from andalusite because of the large amount of overstepping required. Such a relationship was observed, however, in one thin section from the Planet-Huronian Lake area. Uncertainty of fibrolite stability implies a diagnostic value in terms of accurate pressure-temperature estimation less than that of sillimanite itself. Nevertheless, the occurrence of Assemblage 11 (Table 1, Fig. 6) with andalusite as the main aluminum silicate polymorph and staurolite as smaller scattered 'droplets' suggests that these rocks have undergone metamorphism close to reaction (3) conditions and adjacent to the andalusite-sillimanite transition, that is at 550°C and 3 kb (Fig. 3).

Some 4 km west of this locality, quartz monzonite and pegmatite become increasingly abundant and commonly are the predominant rocks. Only at one locality was a significant pelitic assemblage found, namely biotite – garnet – cordierite – sillimanite – 'melt' (cf. Assemblage 13, Table 1) in which the 'melt' is of quartz monzonitic composition. The presence of 'melt' suggests temperatures of at least 660°C at a pressure of 3.5 kb. Assemblage 13 indicates that P-T conditions may have reached slightly beyond those of reaction (4) (Fig. 3).

SHEBANDOWAN – DRIFT LAKE AREA

This area (Fig. 7) was mapped by Morin (1973). A more detailed petrographic study of the metasediments and related migmatites as well as a microprobe study on the chemical zoning of garnets along logging road GLP 511 about 2 km east of Shebandowan was carried out by Birk (1971). A partial section of the Quetico metasedimentary sequence northwards from the fault-bounded Shebandowan metavolcanic belt comprises mainly medium bedded wacke-mudstone units in which metamorphic recrystallization was complete enough to erase grain gradations which, in low grade metamorphic rocks elsewhere, provide facing directions.

The rocks on the north side of the Postans Fault contain biotite but neither chlorite nor sericite, which suggests that in this area the fault has eliminated the lower grade part of the metasedimentary sequence. Birk (1971) observed garnet with biotite in thin sections from several localities. In one sample north of the fault he noted the assemblage staurolite – biotite – garnet (Assemblage 10, Fig. 7) and not far beyond this, the assemblage biotite – garnet – cordierite – staurolite (Assemblage 16, Fig. 7, Table 1). From the texture it appears that cordierite has formed by a reaction consuming staurolite; the lack of aluminum silicate precludes reaction (3) (Fig. 3) in this instance. One possible reaction,

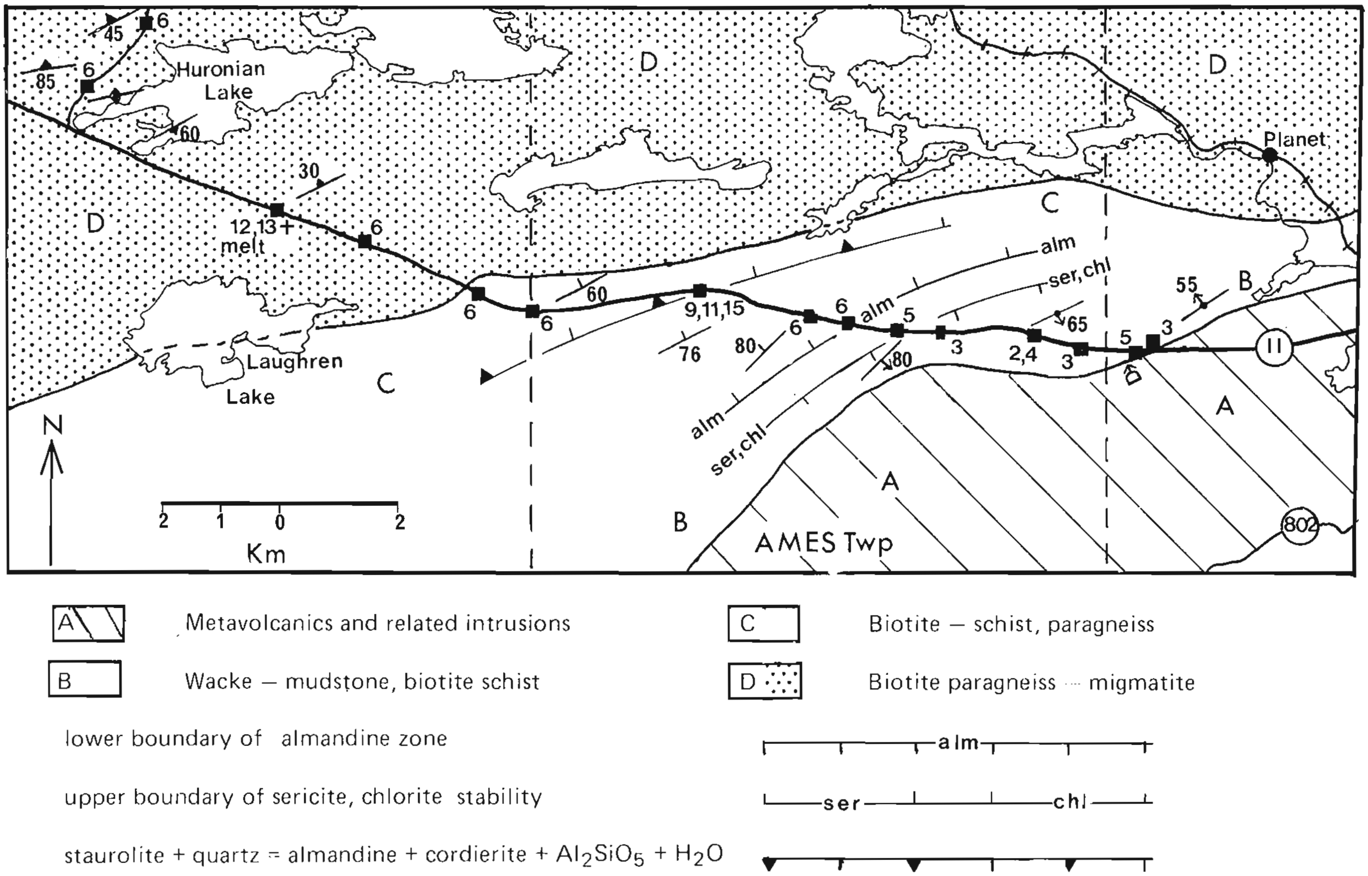

Figure 6. Geology of Planet-Huronian Lake area with approximate metamorphic zones and isograds. Locations of pelitic mineral assemblages are numbered as in Table 1.

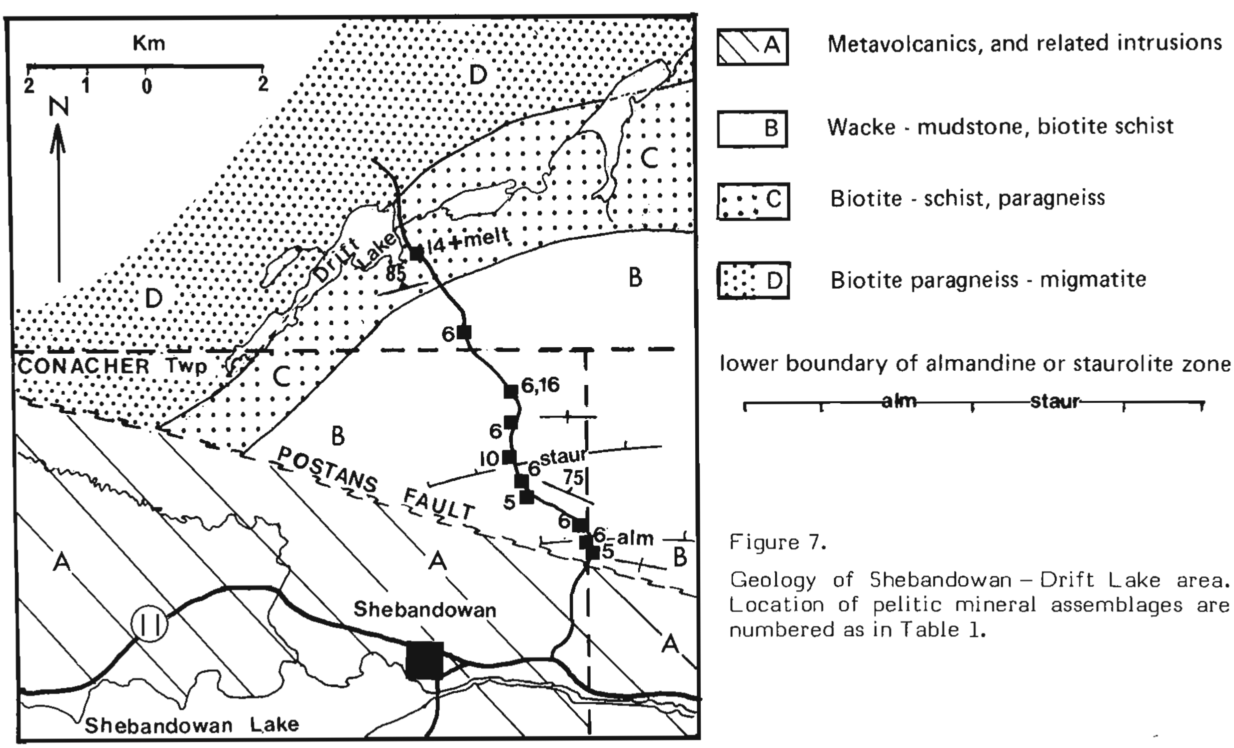

Figure 7.

Geology of Shebandowan – Drift Lake area. Location of pelitic mineral assemblages are numbered as in Table 1.

which forms the assemblage cordierite – garnet – biotite indirectly from staurolite, was indicated by Hess (1969) as being feasible at low pressures:

$$\text{chlorite} + \text{muscovite} + \text{quartz} = \text{garnet} + \text{cordierite} + \text{biotite} + \text{water} \qquad \text{(R.5)}$$

Near Drift Lake the metasediments are represented by biotite paragneiss with substantial 'melt' component. At one locality the assemblage comprises biotite – sillimanite – 'melt' (Assemblage 14 +- melt, Fig. 6) in which the 'melt' is coarse grained quartz, plagioclase, and microcline. Late randomly oriented, coarse muscovite overprints the biotite-sillimanite knots, suggesting the retrograde reation:

$$\text{muscovite} + \text{Na feldspar} + \text{quartz} = \text{K-feldspar} + \text{sillimanite} + \text{melt} \qquad \text{(R.6)}$$

which takes place at about 700°C at a pressure of 4 kb (Fig. 3).

JELLICOE AREA

A similar pattern of progressive metamorphism of typical turbidites is present in the Jellicoe area (Fig. 8). Mackasey et al. (1976) demonstrated that no major break is present between the metavolcanic-metasedimentary Wabigoon Belt to the north and the metasedimentary Quetico Belt to the south. For purposes of this paper the southern mafic volcanic unit of the Wabigoon Belt which trends eastwards along the southern boundary of Legault and Colter townships may be considered to be the boundary between the two belts. The wacke-mudstone beds south of this mafic unit commonly display sedimentary structures such as graded bedding, channelling, and crossbedding. In places facing directions alternate, suggesting that the steeply dipping metasediments are isoclinally folded, but the absence of marker horizons precludes detailed structural analysis. In the metasediments north of the mafic volcanic unit are sericite – chlorite assemblages; to the south biotite stabilizes and sericite disappears.

Pelitic rocks near the northern boundary of the Quetico Belt contain chlorite with biotite. Chlorite disappears southwards; about 3 km farther south garnet appears. Within the next 4 km staurolite is present with garnet and biotite (Assemblage 10, Fig. 8; Table 1). Outcrop is poor, and pelitic assemblages are few in the Jellicoe area, but from observations made in the immediate area to the west the staurolite "in" line has been inferred to trend east-northeast. A similar sequence of mineral assemblages was noted by Macdonald (1944) in the Kenogamisis River area which lies east of the Jellicoe area.

Andalusite and cordierite appear about 4 km or more south of the staurolite isograd (Fig. 8); they do not occur together. Cordierite poikiloblasts in Assemblage 16 (Table 1) contain small staurolite remnants, suggesting staurolite was consumed in the formation of cordierite, as was the case in the Shebandowan-Drift Lake area. Andalusite in Assemblage 9 also contains minor staurolite droplets where muscovite as an additional phase is present, suggesting proximity to the reaction (1) isograd which coincides approximately with the contact of a pegmatite and quartz monzonite body (Fig. 8). In the Jellicoe area the distance from very low grade assemblages to the location of this isograd is substantially greater than in the Crooked Pine Lake area.

GEORGIA LAKE AREA

The Georgia Lake area mapped by Pye (1965) is largely underlain by typical turbidite metasediments of the Quetico Belt which are partly covered by the Keweenawan diabase sheets. Where sedimentary structures are preserved the

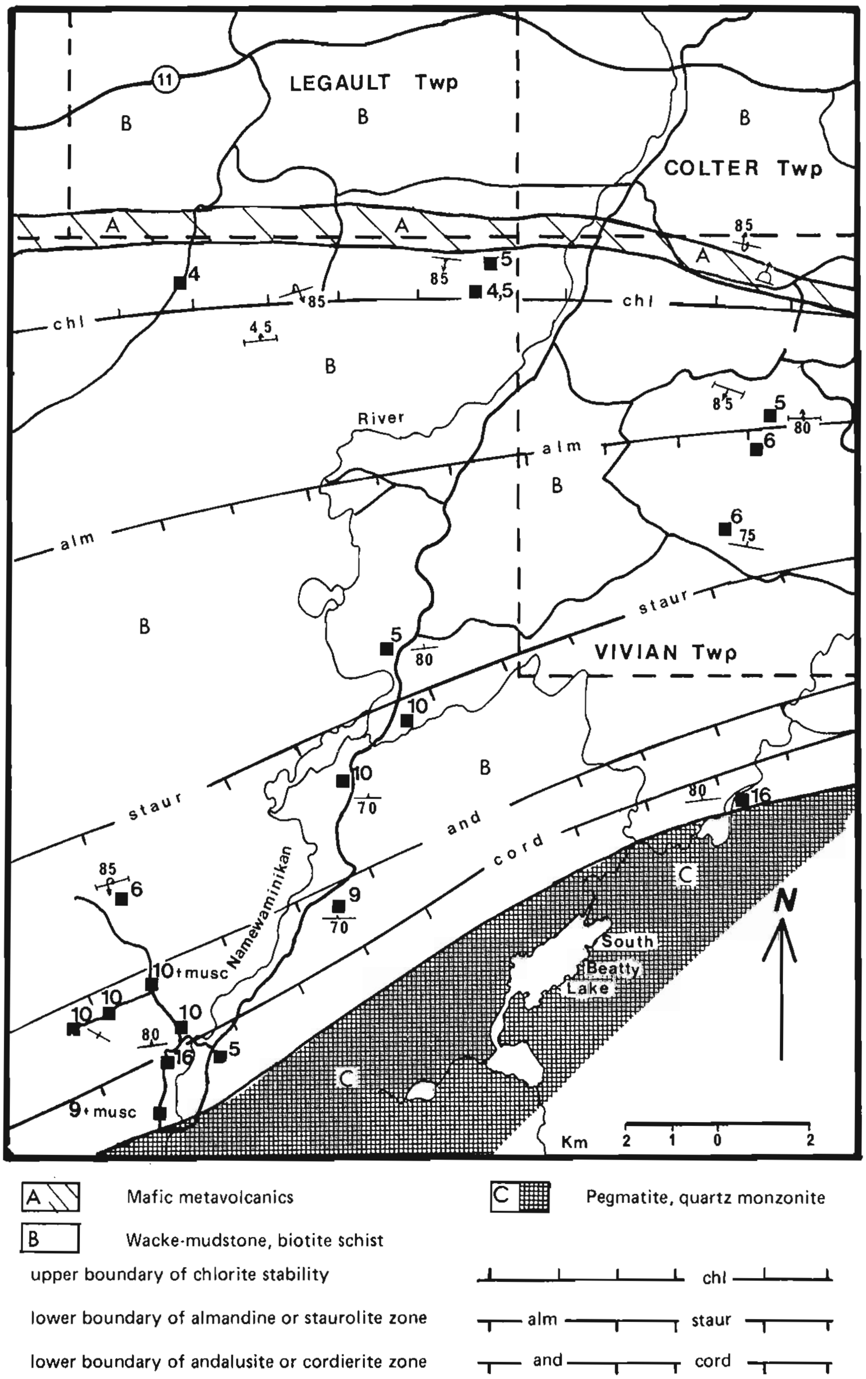

Figure 8. Geology of Jellicoe area with approximate metamorphic zones. Location of pelitic mineral assemblages are numbered as in Table 1.

metasediments face northwards and generally trend northeast to east. The least metamorphosed northern segment contains biotite – garnet assemblages and to the south staurolite and cordierite accompany biotite and garnet. Towards the southern boundary, sheets and dykes of quartz monzonite and pegmatite are progressively more abundant and in the relatively homogeneous quartz monzonite only minor blocks and inclusions of biotite paragneiss remain.

FLANDERS LAKE AREA

The Flanders Lake area, northeast of Manitouwadge, mapped by Milne (1964), is wholly within the migmatized portion of the Quetico Belt. Milne (1964) noted a general increase in metamorphic grade from south to north. Staurolite-bearing migmatized metasediments were found in the south and cordierite is locally present in both the paragneiss and pegmatitic mobilizate portions of the migmatites throughout much of the area. Locally, migmatized paragneiss containing biotite – muscovite – cordierite – sillimanite and biotite – garnet – cordierite – sillimanite assemblages show a close similarity to rocks observed in the Crooked Pine Lake area about 400 km to the west.

In the area adjoining the northwest corner of Flanders Lake area on the west, Coates (1968) recorded the assemblage biotite – garnet – cordierite – hypersthene in biotite paragneiss; the hypersthene is partly retrograded to amphibole and Fe-oxide. This assemblage indicates that granulite facies conditions were imposed on the central regions of the Quetico Belt.

SUMMARY

Sections from the north and south margins and the central part of the Quetico Belt show strong similarities in lithology and regional metamorphic effects. The medium bedded wacke-mudstone turbidites are pelitic enough to illustrate the general nature of the metamorphism from extremely low grade assemblages through a sequence of fairly low pressure, medium grade assemblages characterized by the presence of staurolite, andalusite, and garnet, into high grade assemblages with cordierite, sillimanite, and granitic melt phases.

A tentative pressure-temperature path of metamorphism in the Quetico Belt is outlined in Figure 3. The low grade metamorphism is not well defined because of the lack of compositional data for minerals and of accurate determinations of phase equilibria. Observations in Crooked Pine Lake area of the sequence biotite + garnet + andalusite, andalusite "out", cordierite + garnet + sillimanite over a short distance, suggest temperatures of about 550°C at a pressure of 3 kb, culminating to the south in anatexis of the granitic component of the rocks. The partial retrograding, in the Shebandowan-Drift Lake area, of the assemblage sillimanite – K-feldspar – biotite to yield muscovite appears to have been facilitated by reaction with granitic melt. Because of this and the lack of K-feldspar in the paragneiss the path of increasing metamorphism has been tentatively placed to intersect the granite solidus on the low temperature side of the muscovite breakdown curve. The observed assemblage garnet – cordierite – sillimanite – biotite – melt in the Planet-Huronian Lake area suggests temperatures there were in excess of 720°C at a pressure of about 4.2 kb, as indicated by reaction (4) (Fig. 3).

The highest temperature achieved during metamorphism over most of the migmatized terrane is constrained by the reaction curve for the breakdown of biotite + quartz because of the common occurrence of this association in the biotite paragneiss layers, blocks, and xenoliths within the most completely migmatized portions of the area examined. Locally, however, temperatures may have exceeded those of the reaction

$$\text{biotite + garnet + Na feldspar + quartz = K-feldspar + cordierite + hypersthene + melt} \quad \text{(R.7)}$$

reaching 765°C or more at a pressure of 4.5 kb (Fig. 3) at the climax of metamorphism.

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METAMORPHIC AND TECTONIC EVOLUTION OF THE UCHI-ENGLISH RIVER SUBPROVINCE

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Thurston, P.C. and Breaks, F.W., Metamorphic and tectonic evolution of the Uchi-English River subprovince; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 49-62, 1978.

Abstract

The Uchi subprovince, a subdivision of the Superior Province of the Canadian Shield, is a belt of metavolcanic and granitic rocks with minor metasedimentary rocks. These grade transitionally southward into the English River subprovince, a "gneissic belt" composed of metasedimentary and plutonic rocks. The lithologic transition between the two subprovinces is marked in part by an east-trending zone of cataclasis. The Uchi subprovince is characterized by low pressure subgreenschist to greenschist facies assemblages, whereas the English River subprovince is characterized by slightly higher pressure amphibolite to granulite facies assemblages. The metamorphic zones recognized are low grade: 1) chlorite-biotite zone; medium grade: 2) staurolite-chlorite-biotite zone; 3) sillimanite-muscovite zone; high grade: 4) sillimanite-K feldspar zone, 5) cordierite-almandine-K feldspar zone, and 6) low pressure granulite facies. The lower grade zones occur in the Uchi subprovince and the higher grade zones, south of the cataclastic zone within the English River subprovince. The granulite facies zones do not exceed 3080 Ma in age.

The metamorphic zones suggest the presence of a thermal anticline within the English River subprovince. Metamorphic zonation parallels both the east-trending cataclastic zone and subsidiary north-trending strike slip faults branching off the main zone. The faults are interpreted as thrust faults from the presence of metamorphic zonation, hematite along the fault planes, and mantled gneiss domes. Gneiss domes north of the fault and both the Bee Lake and Red Lake metavolcanic belts are interpreted in part as allochthonous fragments emplaced by nappe tectonics.

Résumé

La sous-province d'Uchi, qui est une subdivision de la province du lac Supérieur, dans le Bouclier canadien, est une zone de roches granitiques et métavolcaniques accompagnées de quelque couches métasédimentaires. Celle-ci passe progressivement vers le sud, dans la sous-province de English River, à une "zone gneissique" composée de roches métasédimentaires et plutoniques. La transition lithologique entre les deux sous-provinces est marquée en partie par une zone de cataclase d'orientation est. La sous-province d'Uchi est caractérisée par des assemblages du faciès des sous-schistes verts et des schistes verts (faciès de basse pression), tandis que la sous-province de English River est caractérisée par des assemblages minéraux du faciès des amphibolites et des granulites (pression légèrement plus élevée). Les zones métamorphiques identifiées sont des zones de métamorphisme de faible intensité: 1) zone à chlorite et biotite; d'intensité moyenne: 2) zone à staurolite-chlorite-biotite, 3) zone à sillimanite et muscovite; d'intensité élevée: 4) zone à sillimanite et K-feldspath, 5) zone à cordiérite, almandine et K-feldspath, et 6) faciès des granulites (basse pression). Les zones de moindre intensité se trouvent dans la sous-province d'Uchi, et les zones de plus forte intensité au sud de la zone de cataclase, à l'intérieur de la sous-province de English River. Les zones appartenant au faciès des granulites n'ont pas plus de 3080 millions d'années.

Les zones métamorphiques semblent indiquer la présence d'un anticlinal thermique dans la sous-province de English River. Les zones métamorphiques sont parallèles à la zone de cataclase d'orientation est, et aux failles de décrochement horizontal d'orientation nord qui lui sont associées et ont leur origine dans la zone principale. On interprète les failles comme étant des failles de chevauchement en raison de l'existence d'une zonalité métamorphique, de la présence d'hématite le long des plans de faille et de domes de gneiss formés par remobilisation du manteau. On estime que les domes de gneiss situés aux abords de la zone faillée et les zones métavolcaniques de Bee Lake et Red Lake sont, au moins en partie, des fragments allochtones mis en place par tectonique en nappes.

INTRODUCTION

A metamorphic map of northwestern Ontario north of the Wabigoon Belt was compiled at a scale of 1:1 000 000 from published reports and work in progress by the geological staff of the Ontario Ministry of Natural Resources. Data for large parts of the area are based on 1:250 000 maps (Bennett and Riley, 1968; Thurston and Carter, 1970; Thurston et al., 1974). The compilation has shown analogies with younger terranes which contribute to the understanding of the evolution of gneiss belts and greenstone-granite terrane in the English River and Uchi subprovinces.

GENERAL GEOLOGY

The English River Subprovince (Fig. 1) is a gneissic terrane consisting of a northern supracrustal domain of metasediments and minor metavolcanic rocks. To the south it is migmatized and gradationally superseded by the southern plutonic domain (Breaks et al., 1974, 1975; Breaks and Bond, 1976, 1977a). The metasediments are stratigraphically equivalent to the metavolcanics of the first of three mafic to felsic volcanic cycles in the Confederation Lake area of the Uchi Subprovince (Fig. 1) (Thurston, 1976) and are also probably equivalent to the upper volcanic cycles. The

[1] Ontario Geological Survey, Ministry of Natural Resources

northern boundary of the English River Subprovince is gradational in stratigraphic terms with the Uchi Subprovince to the north, but is marked by the Sydney Lake Fault System (Wilson, 1971; Stone, 1976, 1977). Portions of the southern plutonic domain of the English River Subprovince are metatonalite 3080 Ma in age (Krogh et al., 1976) whereas the metavolcanics of the Uchi Subprovince are 2960 to 2740 Ma in age (Krogh and Davis, 1971; Nunes and Thurston, 1978).

The Uchi Subprovince, north of the English River Subprovince, is a granite-greenstone block. It is dominated by several mafic to felsic volcanic piles. These are interconnected by the distal facies basaltic flows to form a relatively continuous metavolcanic-metasedimentary complex extending from Lake Winnipeg to the Archean-Paleozoic contact west of Hudson Bay.

Northwest of this subprovince lies the Berens River Subprovince (Wilson, 1971) bounded by faults and consisting of plutonic rocks, and orthogneisses and paragneisses containing remnants of metavolcanic belts now metamorphosed to upper amphibolite facies.

The Cross Lake Subprovince, north of the Berens River and Uchi subprovinces, is an east-west trending granite-greenstone terrane. It is dominated by ovoid, largely unmetamorphosed, gregarious batholiths (MacGregor, 1951) separated by thin, synclinally folded sequences of supracrustal rocks. The subprovince is bordered on the north by a fault-bounded trough termed the Kenyon Structure (Ayres, 1971) filled with high rank psammitic metasediments (Thurston et al., 1974).

VARIATION IN METAMORPHIC GRADE AND STYLE

Metamorphic domains have been divided into 5 zones (Winkler, 1974): very low grade, low grade, medium grade, high grade, and granulite facies (regional hypersthene zone). (Table 1.) The very low grade domain is not well-documented in the Confederation Lake belt. Minor areas characterized by prehnite-pumpellyite and extensive areas of lower greenschist facies, are found in the central portion of the Confederation Lake metavolcanic-metasedimentary belt. Very little recrystallization has taken place and features such as relict

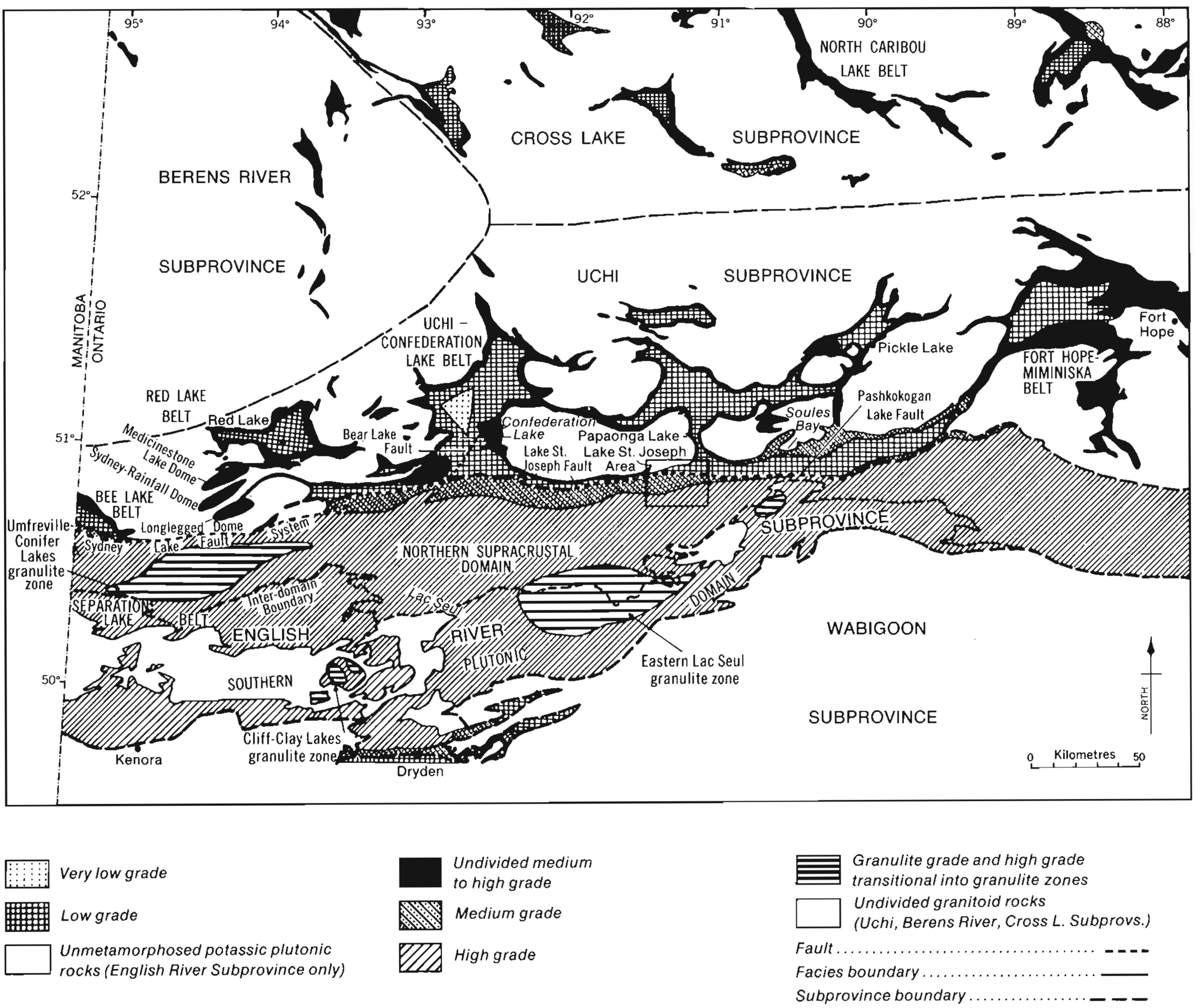

Figure 1. Geological subprovinces and the generalized distribution of metamorphic zones in northwestern Ontario.

pyroxenes in mafic flows and relict snowflake texture in rhyolitic rocks are well preserved (Thurston, 1978). Low grade domains range from mid-greenschist (biotite zone) to the greenschist-amphibolite facies transition. The latter, marked by extensive recrystallization, is confined to the larger metavolcanic-metasedimentary belts. Medium grade domains are characterized by assemblages of the lower to mid-amphibolite facies; high grade domains by mid- to upper amphibolite facies. The transition from medium to high grade is marked by the reaction:

$$\text{muscovite} + \text{quartz} = \text{K feldspar} + Al_2SiO_5 \text{ polymorph} + H_2O/\text{melt} \qquad (R.1)$$

Scattered areas of granulite facies metamorphic assemblages are found in the southern portion of the English River Subprovince. Assemblages in aluminous metasediments indicate that, in general, metamorphism was of the low to intermediate pressure type typical of many shield areas (Binns et al., 1976), shown by the presence of andalusite at moderate grades and sillimanite at higher grades. Kyanite is found only in a few areas where tectonic over-pressures are likely to have been a factor, such as near the faulted northern boundary of the Berens River Subprovince (Wilson, 1971).

There is, however, a tendency to progress from medium grade assemblages containing cordierite without almandine in the metavolcanic-metasedimentary belts to cordierite with almandine in the northern part of the English River Subprovince, suggesting an increase in relative pressure associated with the higher grade (Winkler, 1974, p. 91).

Metamorphic and deformational histories of the English River and Uchi subprovinces are similar (Thurston, 1978; Breaks and Bond, in prep.) involving a maximum of five metamorphic events (Table 2). This interpretation is similar to that derived by McRitchie and Weber (1971) for the Manigotagan gneisses.

Table 1

Metamorphic assemblages in the Cross Lake, Uchi, and English River subprovinces, Northwestern Ontario.

Mafic metavolcanics	Very Low Grade	Low Grade	Medium Grade	High Grade
Chlorite	—	—		
Epidote-Clinozoisite	—	—	— ---	
Actinolite	—	—		
Pale hornblende			—	
Dark hornblende			---	—
Stilpnomelane	—			
Clinopyroxene				---
Orthopyroxene				---
Biotite		---	---	---
Na plagioclase	—	—	—	
Ca plagioclase		--- —	—	—
Iron-Formations				
Chlorite	—	— ---	---	
Minnesotaite	—	— ---	---	
Stilpnomelane	---			
Pelites and semipelites				
White mica	—	—	—	
Chlorite	—	—	— ---	
Biotite		--- —	—	—
Andalusite			—	—
Sillimanite			--- —	— ---
Cordierite			—	—
Almandine			- —	—
Chloritoid	—	—		
Staurolite			— ---	
Na-plagioclase	—	—		
Ca-plagioclase			—	—
Felsic volcanics				
White mica	—	—	—	—
Chlorite	—	—		
Biotite		—	—	—
Pumpellyite	---			
Stilpnomelane	---			
Epidote-clinozoisite	---	---	---	
Actinolite	---	---		
Hornblende			---	---
Almandine			---	---
Na-plagioclase	—	—		
Ca-plagioclase			—	—

Table 2

Metamorphic and deformational events in the Uchi and English River subprovinces (After McRitchie and Weber, 1971)

		S_0	Original sedimentary and volcanic fabric.
D_1			Isoclinal folds in volcanic sequences and nappes in the Red L.-Bee L. areas.
	M_1	S_1	Development of planar fabric preserved as inclusion trains in staurolite, biotite, and almandine, and andalusite.
	M_1A		Main regional metamorphic event. Development chlorite, biotite, hornblende, muscovite, cordierite, almandine, sillimanite. Migmatization of metasediments in English River Subprovince probably commenced during this event.
D_2			Regional folding, rotation of M_1 porphyroblasts associated with emplacement of granitic intrusions in Uchi volcanic sequence.
	M_2	S_2	Matrix coarsening and development of main axial plane schistosity of biotite and muscovite parallel to D_2 folds. Further migmatization of metasediments in English River Subprovince; minor volumes of mobilizate controlled by axial surfaces of mesoscopic D_2 folds.
D_3			Large scale S folds.
	M_3	S_3	Muscovite parallel to D_3 axial planes, pinitization of cordierite and andalusite.
D_4		S_4	Late stage development of mylonite on strike slip faults.
	M_4		Retrograde muscovite and chlorite in shear zones of D_4.
D_5		S_5	Late transcurrent faulting, Bear Lake Fault.
	M_5		Minor recrystallization associated with D_5.

The earliest metamorphism, M_1, resulted in the formation of porphyroblasts of staurolite, andalusite, biotite, and almandine garnet. These porphyroblasts with inclusion trails of biotite, quartz and feldspar, have been rotated during deformation D_2. Metamorphism M_2 resulted in the coarsening of the matrix, particularly of metasediments, and the growth of biotite and muscovite grains parallel to the axial plane of D_2 folds. Metamorphism M_3 was largely retrograde, as recorded in the pinitization of cordierite, sericitization of andalusite, and alteration of amphiboles to chlorite. Metamorphism M_4 was essentially local, associated with regeneration of the Sydney Lake Fault, separating the English River and Uchi subprovinces, and generation of various major strike-slip fault components in the metavolcanic terrane in the Lake St. Joseph area. The main features of this event, as observed immediately adjacent to the faults, are recrystallization of muscovite and chlorite, formation of hematite, carbonate, epidote and minor pyrite, and diaphthoretic replacement of M_2 textures.

Transcurrent faults such as the Bear Lake Fault are much later than the Sydney Lake Fault and its various branches (Fig. 1). Late recrystallization associated with these faults may mark a fifth metamorphic event, M_5 (Thurston, 1978).

METAMORPHISM IN THE ENGLISH RIVER AND UCHI SUBPROVINCES

Northern Supracrustal Domain

Several workers have documented a progressive increase in metamorphic grade coinciding with transition from the metavolcanic-metasedimentary successions of the Uchi Subprovince southwards into the northern metasedimentary domain of the English River Subprovince (McRitchie and Weber, 1971; Jones, 1973; Dwibedi, 1966). The progressive regional metamorphic patterns are commonly displaced by extensive postmetamorphic fault systems that appear to have been initially guided by the stratigraphic interface between the two subprovinces. For example, the Sydney Lake Fault System currently under investigation by Stone (1977) marks an abrupt "change" from medium grade, low pressure assemblages to high grade migmatitic rocks. Similarly, along the Pashkokogan Lake Fault System southeast of Lake St. Joseph, low grade mafic metavolcanics are in sharp contact with metatexitic metasedimentary migmatite. Progressive metamorphic zones, apparent in the Lake St. Joseph-Papaonga Lake area (Fig. 2), coupled with the data of McRitchie and Weber (1971) indicate that the southerly increasing metamorphic grade from the Uchi Subprovince into the English River Subprovince is a widespread phenomenon involving at least 300 km of the subprovince interface.

Five major metamorphic zones, delineated in the Lake St. Joseph area (Table 3) and based mainly upon assemblages recorded in pelitic and wacke compositions are:

LOW GRADE

(1) chlorite-biotite zone

MEDIUM GRADE

(2) staurolite-chlorite-biotite zone

(3) sillimanite-muscovite zone

HIGH GRADE

(4) sillimanite-K feldspar zone

(5) cordierite-almandine-K feldspar zone

A plausible reaction for formation of M_2 staurolites (Table 2) may be similar to that experimentally determined by Hoscheck (1969):

$$\text{chlorite + muscovite = staurolite + biotite + quartz + vapour} \quad (R.2)$$

Chlorite and muscovite are widespread in pelitic rocks of the chlorite-biotite zone of the low grade metamorphism which immediately adjoins the staurolite zone to the north. Chloritoid does not appear to be a significant phase in the chlorite-biotite zone. Chlorite, biotite, and muscovite also represent integral phases in the S_2 foliation of the staurolite zone.

Cordierite, aligned within M_2 foliation surfaces and coexisting with sillimanite, muscovite, almandine, and biotite, appears as porphyroblasts, prior to elimination of M_2

staurolite with increasing metamorphic grade. The following experimentally determined reaction (Hirschberg and Winkler, 1968) appears to be applicable:

chlorite + muscovite + quartz = cordierite + biotite + sillimanite + vapour (R.3)

Relicts of staurolite occur in the centres of some cordierites, but the relationship of these minerals is not clear. It is also uncertain whether M_2 staurolites developed within the sillimanite or within the andalusite P-T field of stability. These staurolites have not been noted with any of the Al_2SiO_5 minerals although in the Soules Bay area of Lake St. Joseph (Fig. 1) to the east, coexisting staurolite and andalusite porphyroblasts have been observed (Sage and Breaks, 1976). It seems plausible that the path of metamorphism breached the andalusite-sillimanite boundary within the staurolite zone prior to the first appearance of cordierite. It should be mentioned that a zone of high level cataclastic deformation (Lake St. Joseph Fault Zone) crosses the staurolite zone. This zone of brittle failure postdates the youngest Archean granitoid phases (leucocratic quartz monzonite) and regional metamorphism, although no significant change in metamorphic grade is apparent across the fault. The main effect of this fault was to facilitate H_2O ingress into wall rocks causing retrogression of M_2 metamorphic assemblages, as exemplified by complete replacement of staurolite by chlorite and muscovite. Obliteration of primary structures in the high grade zone is related to anatexis and pervasive matrix coarsening.

Mineral assemblages corresponding to each of the zones listed in Table 3 may be explained in terms of experimentally and/or petrographically validated metamorphic reactions. The observed increase in grade corroborates a transition from nonmigmatized metasedimentary assemblages in the medium grade zones to an orderly succession of migmatitic stages at high grade (Breaks and Bond, 1977a).

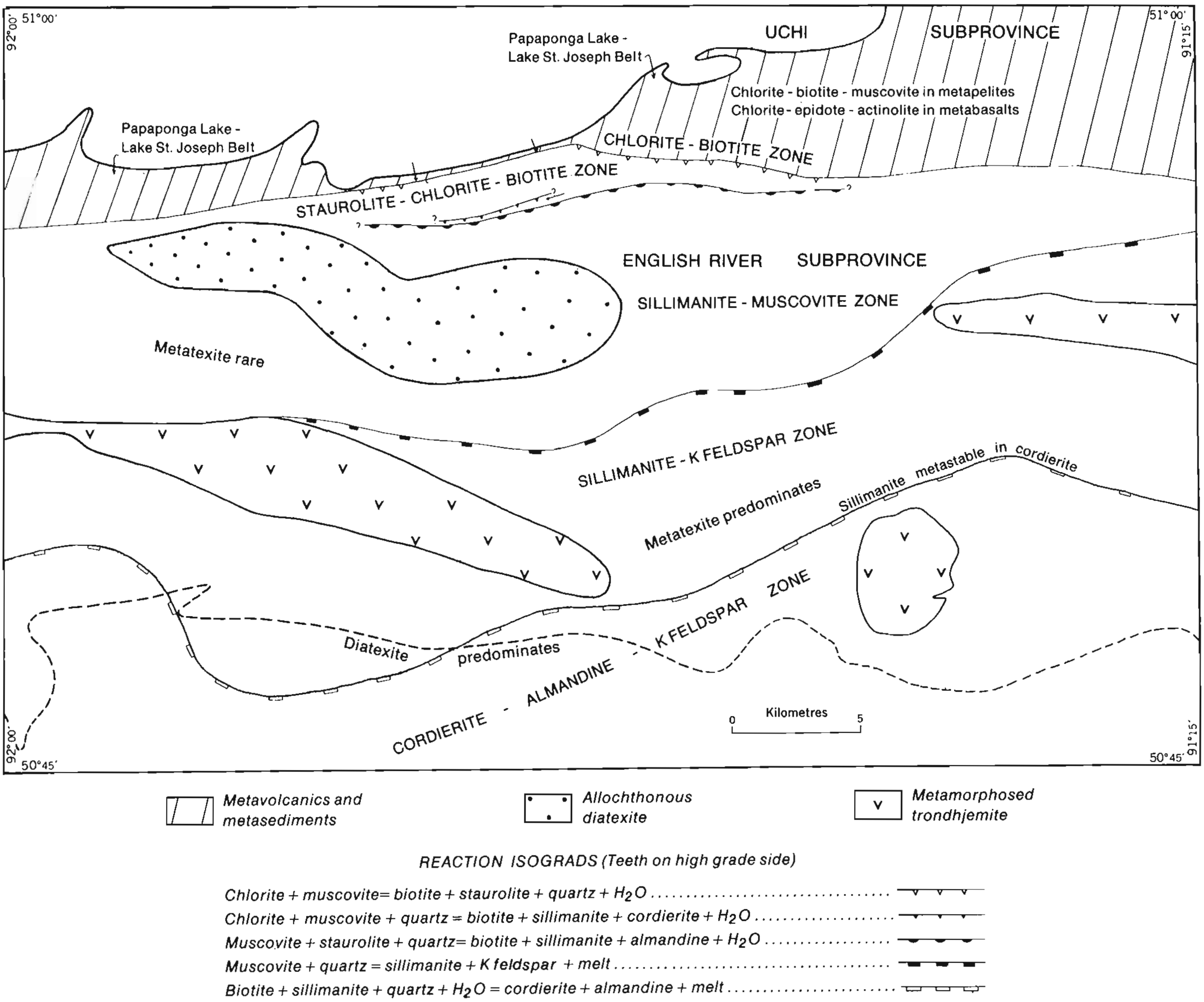

Figure 2. Distribution of metamorphic zones, isograds, and metasedimentary migmatite stages at Uchi Subprovince-English River Subprovince interface, Lake St. Joseph area.

Table 3

Metamorphic mineral assemblages from the Lake St. Joseph and eastern Lac Seul areas[1]

ZONE	ASSEMBLAGE	ROCK TYPE
	Low Grade	
Chlorite-Biotite	Chl – Mu – Bio – Ep ± Gf	W
	Mu – Chl ± Gf – Tour	W
	Bio – Chl – Ep – Act	W
	Bio – Mu – Chl – Gf	P
	Bio – Ep – Mu	W
	Chl – Act – Ep	M
	Act – Hb – Bio	M
	Medium Grade	
Staurolite-Chlorite-Biotite	Mu – Chl – Bio	W
	Bio – Mu	W
	St – Bio – Chl – Mu ± Chl, Mu	P
	Mu – Bio – St – Alm ± (Alm)	P
	Bio – St – Mu ± Chl, Mu	P
	Bio – Chl – Mu ± (St, And)	P
	Bio – Sill – Alm ± Gf ± (Alm)	P
	Sill – Bio – Mu ± Chl	P
	Bio – Alm – Sill ± Tour ± Gf ± (St)	P
	Bio – Cord – Sill – Alm ± Mu	P
	Bio – Mu – Sill ± Alm ± Tour ± (St)	P
	Trem – Bio	UM
	Bio – Cord – Alm ± (St)	P
	Medium Grade (cont.)	
Sillimanite-Muscovite	Bio – Mu	W
	Bio – Mu – Alm	W
	Bio – Sill – Mu – Alm	P
	Bio – Mu – Sill	P
	Bio – Sill – Mu ± Tour ± (St)	P
	Bio – Sill – Alm – Mu ± Gf	P
	Bio – Cord – Sill ± Alm ± Tour	W
	Bio – Sill ± (Mu)	P
	Bio – Cord – Alm – Mu	W
	Bio – Cord	W
	High Grade	
Sillimanite-K-Feldspar	Bio – Alm	W
	Bio – Alm – Cord	P
	Bio – Kspar – Cord – Sill	P
	Bio – Cord – Sill – Alm ± Kspar	P
	Bio – Sill – Cord – Alm	P
Almandine-Cordierite-K feldspar	Bio – Alm	W
	Bio – Alm ± Mu	W
	Bio – Cord – Alm ± (Sill)	P
	Bio – Cord ± Alm ± Kspar ± (Sill)	P
	Kspar – Bio – Cord	P
Low pressure Granulite Zone	Bio – Opx – Cpx – Hbl ± Alm ± Qtz	M
	Opx – Cpx – Hbl ± Alm	M
	Bio – Cord – Alm – Kspar ± (Sill)	P
	Opx – Cord – Bio – Hbl	I

Abbreviations

Act	=	actinolite	Mu	=	muscovite
Alm	=	almandine garnet	Sill	=	sillimanite
And	=	andalusite	St	=	staurolite
Bio	=	biotite	Tour	=	tourmaline
Chl	=	chlorite	Trem	=	tremolite
Cord	=	cordierite	I	=	intermediate granitoid rocks
Ep	=	epidote/clinozoisite	M	=	mafic metavolcanics, mafic dykes
Gf	=	graphite	P	=	pelitic metasediments
Hb	=	hornblende	UM	=	ultramafic rocks
Kspar	=	K feldspar	W	=	wacke metasediments

Minerals enclosed by rounded brackets () represent porphyroblastic M_1 relict phases exhibiting preserved internal foliation. Those enclosed by squared brackets represent M_3 diaphthoritic mineral phases.

[1] quartz and plagioclase common to all assemblages except those of UM bulk composition.

Within the medium grade zones, sedimentary bedding structures and framework grains in wackes are generally not completely obliterated due to low P-T conditions. Layers, boudins and rootless intrafolial folds of hydrothermal mobilizate (quartz veins) are, however, evident in the transitional area between the staurolite-chlorite-biotite and sillimanite-muscovite zones. The transition from medium to high grade metamorphism corresponds to disappearance of M_2 muscovite and entry into P-T conditions conducive to widespread development of in situ anatectic melts (metatexis). The metatexitic stage of migmatization generally corresponds to the sillimanite-K feldspar zone. Stromatically disposed leucosomes of granitoid to pegmatoid material are now predominant. Dykes and discordant masses of tourmaline-muscovite mobilizate derived from this zone at depth invaded the lower grade sillimanite-muscovite zone to the north (Fig. 2).

Near the cordierite-almandine-K feldspar isograd the metatexite zone imperceptibly yields to higher grade conditions promoting advanced stage fusion (diatexis) through the reaction:

biotite + sillimanite + quartz + vapour =
cordierite + almandine + melt (R.4)

On the high grade side of the isograd, metastable sillimanite is widely preserved in the cores of cordierite grains. To the south the assemblage biotite – K feldspar – cordierite – almandine is common in remnant metapelitic layers. In general, the proportion of metapelitic paleosome decreases abruptly southwards from the second sillimanite isograd (Fig. 2). A large proportion of this material has been extensively converted to leucosome and melanosome components.

Southern Plutonic Domain

Variation and intensity of regional metamorphic grade within the southern plutonic domain of the English River Subprovince is more enigmatic than that in the north because of the following factors:

(1) a dearth of bulk compositions necessary to record diagnostic metamorphic assemblages;

(2) widespread late Archean invasion by potassic granitoid rocks which has effectively reduced the areal extent of pre-Kenoran metamorphosed, foliated and gneissic granitoid rocks.

Amphibolitic inclusions such as those within the Deception Bay Gneiss Belt north of Sioux Lookout (Breaks and Bond, 1977b) are widespread; however, these bulk compositions are notoriously insensitive to changes in metamorphic conditions under medium to high grade regional metamorphism. Thus, only ubiquitous biotite – hornblende – plagioclase ± quartz ± Fe-oxide assemblages are recorded in these rocks. It appears that P_{Load} was insufficient to stabilize almandine-producing reactions in most of these appropriate mafic compositions, in contrast to the omnipresence of this mineral in the Northern Supracrustal Domain.

According to Winkler (1974, p. 89), load pressures of at least 4 kb[1] at 500°C are generally required initially to stabilize almandine garnet (depending in part upon amount of the spessartine component) in common rock types. Temperatures and vapour pressures appear substantial enough to foster localized anatexis of some amphibolite gneiss units, as in the Deception Bay area. Pods and small masses of white, hololeucocratic and leucocratic quartz-poor trondhjemite and biotite-hornblende quartz diorite exhibit a close, carapace-type relationship with incompletely degraded amphibolitic paleosome. This leucosome megascopically resembles the widespread mobilizate material prevalent in metasedimentary migmatite of the Northern Supracrustal Domain.

Low Pressure Granulite Facies Metamorphism

Several significant areas of the English River Subprovince in Ontario have been affected by granulite facies regional metamorphism. Three areas, intermittently spanning a strike length of 315 km, have been delineated to date (Fig. 1).

1. Umfreville Lake – Conifer Lake zone,
2. Cliff Lake – Clay Lake zone (Westerman, 1978), and,
3. Eastern Lac Seul zone (Urquhart, 1976; Breaks and Bond, 1977b).

Field evidence indicates that the granulite facies overprint took place during the Kenoran tectonic-metamorphic episode. The individual zones vary in scale from single occurrences to rather extensive, ovoid areas, the latter exemplified by the Eastern Lac Seul zone, which is about 85 km long and 30 km wide. The various zones are either superimposed upon the inter-domain boundary or entirely isolated within the northern or southern domains. In the central Manigotagan gneissic belt of Manitoba Trueman et al. (1975) have documented the presence of an orthopyroxene zone of unspecified dimensions.

The following field and mineralogical features of the English River Subprovince granulites together indicate that these rocks originated under P-T conditions pertaining to the low pressure subdivision of Green and Ringwood (1967)[2]:

1) the predominance of felsic rocks with subordinate amounts of intermediate and mafic rocks
2) the presence of cordierite, biotite, and almandine garnet
3) the restriction of orthopyroxene distribution mainly to mafic and intermediate rocks
4) a prograde transition to low pressure granulite conditions, with no evidence of tectonic uplift
5) the absence of associated members of the anorthositic suite and rocks of charnockitic affinities.

Coexistence of cordierite and orthopyroxene is evident, albeit rare, in certain metasedimentary layers. It is usually more common to observe cordierite – almandine – quartz – plagioclase ± K-spar assemblages interlayered with orthopyroxene ± diopside-bearing mafic dykes. Orthopyroxene and/or diopside are more widespread in rocks of mafic and intermediate bulk composition. Field evidence clearly documents prograde reactions leading to elimination of hornblende from amphibolitic horizons of the eastern Lac Seul area (Breaks and Bond, in prep.). This result can be approximated as follows:

hornblende + quartz → orthopyroxene + diopside +
plagioclase +
Fe-oxide (magnetite) + melt (R.5)

Furthermore, development of diopside and/or orthopyroxene-bearing medium- to coarse-grained, leucocratic granitoid partial melt phases in amphibolite and intermediate granitoid bulk compositions attests to physical conditions in which P_{Load} approximates P_{H_2O}. Under these conditions Binns (1969) has indicated experimentally that appearance of a melt phase in quartz in quartz-bearing amphibolite and hornblende-pyroxene gneiss can occur at minimum temperatures of 770°C and pressures exceeding about 2.5 kb.

Hornblende decomposition, however, cannot be considered a widespread process in the ubiquitous wacke and pelitic bulk systems, since this mineral is extremely uncommon in lower grade equivalents due to relatively low calcium contents (usually less than 3.5 per cent). The

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[1] 1 kb = 1 x 10^5 kPa.

[2] also see Lambert and Heier (1968).

appearance of orthopyroxene in some wackes may however, be explicable in terms of the semiquantitative univariant reaction (Grant, 1973, p. 508):

biotite + garnet → orthopyroxene + cordierite + K feldspar + melt (R.6)

The zone of granulite metamorphism in the eastern Lac Seul area, situated about 30 km south of the Lake St. Joseph facies series (Fig. 2) may represent a continuation of the interpreted path of metamorphism outlined in Figure 3, and is reminiscent of the thermal anticlinal model proposed by Richardson (1970). It is possible that progressive removal of fluid phase components via anatectic melts engendered by intersection of the path of metamorphism with melt-producing divariant reactions under high grade conditions, may ultimately yield relatively anhydrous bulk compositions capable of yielding granulite assemblages.

Metamorphic Conditions

Delineation of the path of progressive regional metamorphism southwards from the Uchi Subprovince-English River Subprovince interface (Fig. 3) may be estimated from the aforementioned isograds in conjunction with several important observations:

1. high pressure phases such as kyanite or high temperature high pressure phases as sapphirine are unknown in the English River Subprovince,
2. cordierite is an exceedingly widespread phase in the northern supracrustal domain of the English River Subprovince especially in high grade metamorphic zones,
3. decomposition of staurolites does not coincide with sudden appearance M_{1A} of metatexitic migmatization; stability of staurolite, in other words, does not extend into regions where M_2 muscovite (K feldspar-ilmenite zone) is absent ($P_{Tot} < 4.5$ kb) (Winkler 1974, p. 199, 218),
4. the disappearance of M_2 muscovite in the presence of quartz generally coincides with pervasive metatexitic migmatization of pelitic-wacke assemblages (P-T > 3 kb),
5. appearance of mobilizate-dominant migmatite stages resulting from advanced stage fusion (diatexis) approximately coincides with cordierite-almandine-K feldspar zone,
6. almandine becomes widespread only in the high grade zones suggesting increasing pressure for the facies series represented in the Lake St. Joseph area,
7. localized partial melting of amphibolitic bulk compositions under low pressure granulite conditions indicates P_{Load} approximates P_{H_2O} and P > 2.5 kb, T > 770°C.

Assuming P Total = P_{H_2O} estimates of P-T conditions are as follows:

Medium Grade = 550-650°C, 3-4.5 kb

High Grade = 650-790°C, 3-7.5 kb

Granulite = > 770°C, > 2.5 kb

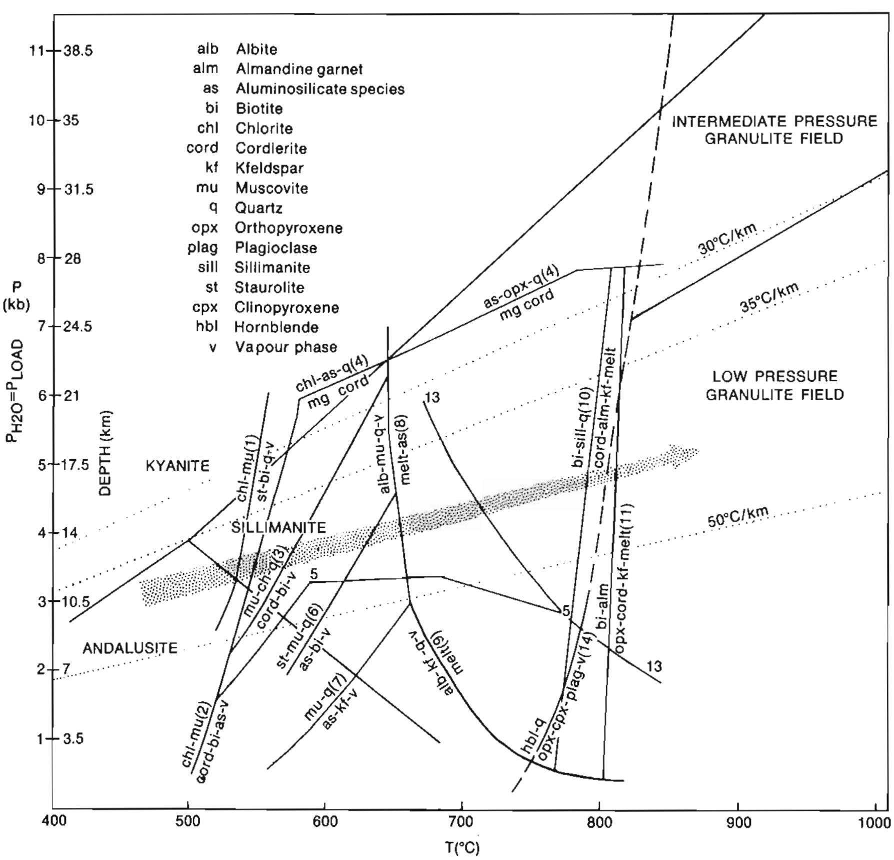

Figure 3. Inferred path of regional metamorphism from Uchi Suprovince transitional into English River Subprovince.

LEGEND

1 = Hoschek (1969),
2 = Hirschberg and Winkler (1968),
3 = Schreyer and Seifert (1969),
4 = upper pressure stability of magnesian cordierite after Schreyer and Seifert (1969) and Newton et al. (1974),
5 = upper pressure stability of Fe cordierite Richardson (1968),
6 = Hoschek (1969),
7 = Althaus et al. (1970),
8 = Storre and Karotke (1971),
9 = after Merrill et al. (1970) and Tuttle and Bowen (1958),
10+11 = Grant (1973), Aluminosilicate system after Holdaway (1971),
12 = Binns (1969),
13 = beginning of melting curve for quartz bearing amphibolite and hornblende pyroxene gneiss after Binns (1969),
14 = boundary between high grade and granulite conditions based on breakdown for common hornblende in presence of quartz (Binns, 1969) and extrapolated beyond 3 kb,
15 = boundary between low and intermediate pressure granulite fields as defined by upper pressure stability of olivine + plagioclase in basaltic compositions with 100 Mg/Mg + Fe^{2+} = 60, Green and Ringwood 1967.

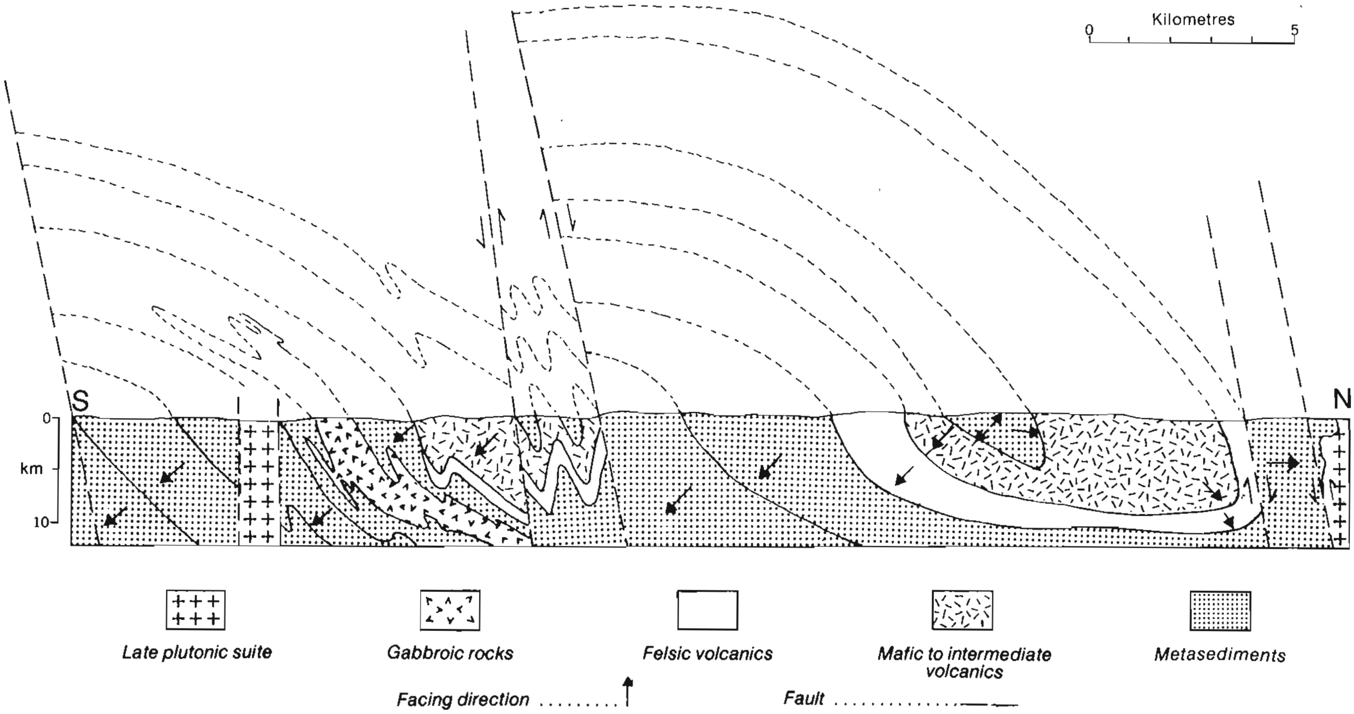

Figure 4. Cross-section of the Bee Lake-Rice Lake volcanic belt. The belt forms the nose of a nappe driven northward by gravity driven deformation.

These conditions are indicative of an average geothermal gradient between 35°C/km and 50°C/km, possessing mineralogical characteristics of the low pressure-intermediate series of Miyashiro (1961, p. 283, 303).

Two styles of metamorphism are represented in the area: static metamorphism in which primary structures (pillows in mafic flows, fragments in pyroclastic rocks) and textures are well preserved and, dynamic metamorphism, characterized by more intense deformation, effecting destruction of original textures and original structures by flattening, elongation, or obliteration by penetrative deformation. Static metamorphism is preferentially distributed in the metavolcanic-metasedimentary belts and may yield gradationally to dynamic metamorphism in the gneiss-belt terranes.

Relationship to Granitic Rocks

Granitic plutons in the English River and Uchi metavolcanic-metasedimentary subprovinces are either forcefully emplaced granitic bodies of mesozonal affinity having relatively wide (1-2 km) contact aureoles of foliated amphibolite facies supracrustal rocks, or diapiric stocks commonly within supracrustal belts surrounded by relatively narrow zones of hornblende and pyroxene hornfels facies rocks.

The low grade metamorphism of the Uchi Subprovince supracrustal belts, except for the contact effects at the margins of the belts, is centred on the Confederation Lake area. In the volcanic areas to the east and west, metamorphic grade gradually increases; the grade is particularly high in the Miminiska-Fort Hope belt.

In general the sodic granitic rocks are associated with the lower grade rocks of the Uchi Subprovince (Thurston, 1978). Saggerson and Turner (1976) note an association of the sodic granitic suite with lower metamorphic grade areas and potassium-rich granites with higher grade areas. This tendency is weakly displayed within the English River Subprovince.

Thin, high grade belts of supracrustal rocks occur between ovoid granitic plutons. Migmatitic structures indicative of anatexis are rare where granitic rocks are in contact with metavolcanic belts.

Relationship to the Gravity Field

The gravity field of the western portion of the Uchi Subprovince consists of gravity highs over supracrustal belts and lows over the surrounding granitic areas. In detail the highs are associated not only with thick sequences of mafic flows but also with the terranes of lowest metamorphic grade within the belt (Barlow et al., 1976).

In the English River Subprovince the northern supracrustal domain as a whole is underlain by a gravity high of about 20 mgals above regional background (West, 1976). The southern granitoid plutonic domain of this subprovince is also associated with a local gravity high amounting to about 15 mgals above regional background (V.K. Gupta, pers. comm.). It is thus evident that regional gravity highs are not strictly correlative with low grade greenstone terranes but may also coincide with high grade gneissic belts, particularly with those areas exhibiting low pressure granulite facies assemblages.

Relationship of Metamorphism to Stratigraphy and Tectonics

The low to very low grade domains occur in the central parts of the metavolcanic-metasedimentary belts, and due to the synclinal nature of most belts, these are also the stratigraphically higher parts of the successions. Proximal facies felsic metavolcanics and metasediments are generally found in lower grade domains. Metamorphic boundaries within the belts appear to crosscut stratigraphic units. Thus the boundary between the low grade terrane of the Uchi Subprovince and the medium grade terrane of the English River Subprovince crosses the mafic metavolcanics of the lowest cycle in the Confederation Lake area at a high angle.

Pre-Cleavage Deformation

The salient features of the region are the transition from the Uchi Subprovince low grade domain to the English River Subprovince high grade domain and the association of the Sydney Lake Fault System with the subprovince boundaries. This metamorphic pattern is probably due to vertical movement of the Sydney Lake Fault System, as postulated by Wilson (1971) and McRitchie and Weber (1971) who also documented a right-lateral component of "almost 10 miles" (16 km) based on dislocation of the contact of a quartz diorite pluton south of Wanipigow Lake in Manitoba. To the east, in the area south of Red Lake, Stone (in press) suggested a minimum right-lateral displacement of 27 km.

Metamorphic data from the vicinity of the Sydney Lake Fault suggest a minimum vertical movement of 4 km. On the basis of structural data, Stone (in press) has inferred vertical movement of 3.3 km. Muscovite-bearing metapelites on the north side of the fault contrast with K feldspar- and andalusite-bearing assemblages on the south side of the structure, 200 m away.

Thrust faulting of large magnitude has been described in the Archean of Rhodesia (Coward et al., 1976a; Coward et al., 1976b; Coward, 1976) where large scale nappes and allochthonous fragments of metavolcanic belts occur 50 to 80 km from their source.

Hematite in the Sydney Lake Fault zone is interpreted to be the result of oxidation of iron-bearing silicates by dissociation of water moving up the fault zone, but down the thermal gradient produced by upthrusted material from the south side of the fault overlying the colder rock on the north side of the fault. (Beach and Fyfe, 1972.)

Trueman et al. (1975) have noted a symmetrical metamorphic zonation in the Manigotagan gneiss, the western extension of the English River Subprovince, which from north to south proceeds from chlorite zone to orthopyroxene granulite to chlorite zone.

Zwart (1974) from work in the Scandinavian Caledonides interpreted stratigraphic thickening and inversion of metamorphic isograds as the result of thrusting and nappe formation. The symmetrical metamorphic pattern at the west end of the English River Subprovince may have developed in a similar manner.

The structure of the Confederation Lake metavolcanic-metasedimentary belt comprises a medial syncline and an eastern north-south striking anticline separated by the Uchi Lake Fault (Thurston, 1976). East of the fault metamorphic grade is low amphibolite facies; west of the fault it is low greenschist facies. The fault is a splay of the Sydney Lake Fault System and, by analogy with the faults and associated fold pattern described by Harris (1970), is thought to be a thrust fault. This fault predates the intrusion of the late granitic rocks in the region.

Thrust faulting may have occurred in areas where the gravity response of the metavolcanic-metasedimentary belt does not correspond closely with the variation in measured stratigraphic thickness. Grant et al. (1965) concluded that the metavolcanics in the Red Lake belt reached a maximum depth of 7.6 km and "that an ancient granitic crust composed partly of sediments must have existed before the formation at its surface of Archean lavas and sediments."

The Rice Lake-Bee Lake metavolcanic-metasedimentary belt in the area of the Sydney Lake Fault System may be an overturned anticline (Stockwell, 1945). Shklanka (1967) suggested that the Bee Lake succession is monoclinal, representing the south limb of the anticline. The authors suggest, however, that it is in part overturned, as indicated by facing directions and by a stratigraphic sequence which progresses, from north to south, from the felsic pyroclastic and metasedimentary top of one cycle through the mafic base to the felsic and metasedimentary upper portion of another cycle. The sequence at some localities yields evidence of overturning with dips of 55° to 65° to the north; remnants of the Bee Lake belt may thus represent an allochthonous fragment. Relationships in the Rice Lake belt to the west are clearer; the area shown in Figure 4 (section E-E' of Weber, 1971, reinterpreted by the authors) comprising part of the eastern portion of the belt is considered to be an antiformal fragment of a nappe.

Interpretation of the Red Lake metavolcanic belt (Fig. 5) as a synformal recumbent anticline, is based upon shape calculations from gravity data (Grant et al. 1965), structural analyses of the belt (Riley, 1975; R.A. Riley, pers. comm., 1977), and the shallow dipping, in places overturned volcanics, in particular at the southern edge of the belt (Ferguson, 1965).

The Red Lake belt forms the nose of a nappe which is infolded into the surrounding granitic rocks and basement. The upper limb of the nappe, extending to the south, is preserved along the flanks of the Rainfall Lake and Sydney Lake gneiss domes (Fig. 1) (Breaks et al. 1974). The nappe hypothesis provides the best explanation of the structural pattern in this area.

Gorman et al. (1978) have suggested that the density inversion produced by mafic volcanic rocks (S.G. 3.0) above a sialic crust (S.G. 2.7-2.9) (Fig. 6) is compensated for by gravitationally driven deformation of the volcanic area into a central syncline with large scale recumbent folds and thrusts parallel to bedding. By use of Talbot's (1974) model studies we suggest that if the interface between the volcanic rocks of the Uchi Subprovince and the metasediments to the south had a dip of greater than 25 degrees northward (Fig. 6) gravity driven deformation would produce a pleurotoid eccentric nappe with the overlying volcanic rocks transported northward.

DISCUSSION

Although the metamorphic path postulated for the English River and Uchi subprovinces is only approximate, because of the lack of mineral composition data, it is probably qualitatively reliable and therefore certain inferences can be drawn. The assemblages present in these subprovinces indicate an approximate pressure of between 3 and 4.7 kb corresponding to a depth of 10.5 to 16 km. A geothermal gradient constructed for points within the Uchi Subprovince and the northern and southern portions of the English River Subprovince is interpreted to lie between 35°C/km and 50°C/km. It is evident that a fairly rapid increase in geothermal gradient occurred. Areas of maximum heat-flow in either the southern or northern domains are now represented by low pressure granulite facies assemblages. The surface pattern of metamorphic zonation can be readily explained by the heat-flow distribution. Zwart (1962) noted geothermal gradients in a similar situation in the order of 150°C/km, therefore the gradients in this area do not appear excessive. Richardson (1970) proposed a petrogenetic model of metamorphic belts in which isotherms in a sedimentary pile are displaced upward under conditions of low pressure metamorphism to form a thermal anticline. This model may be applicable to the Uchi and English River subprovinces.

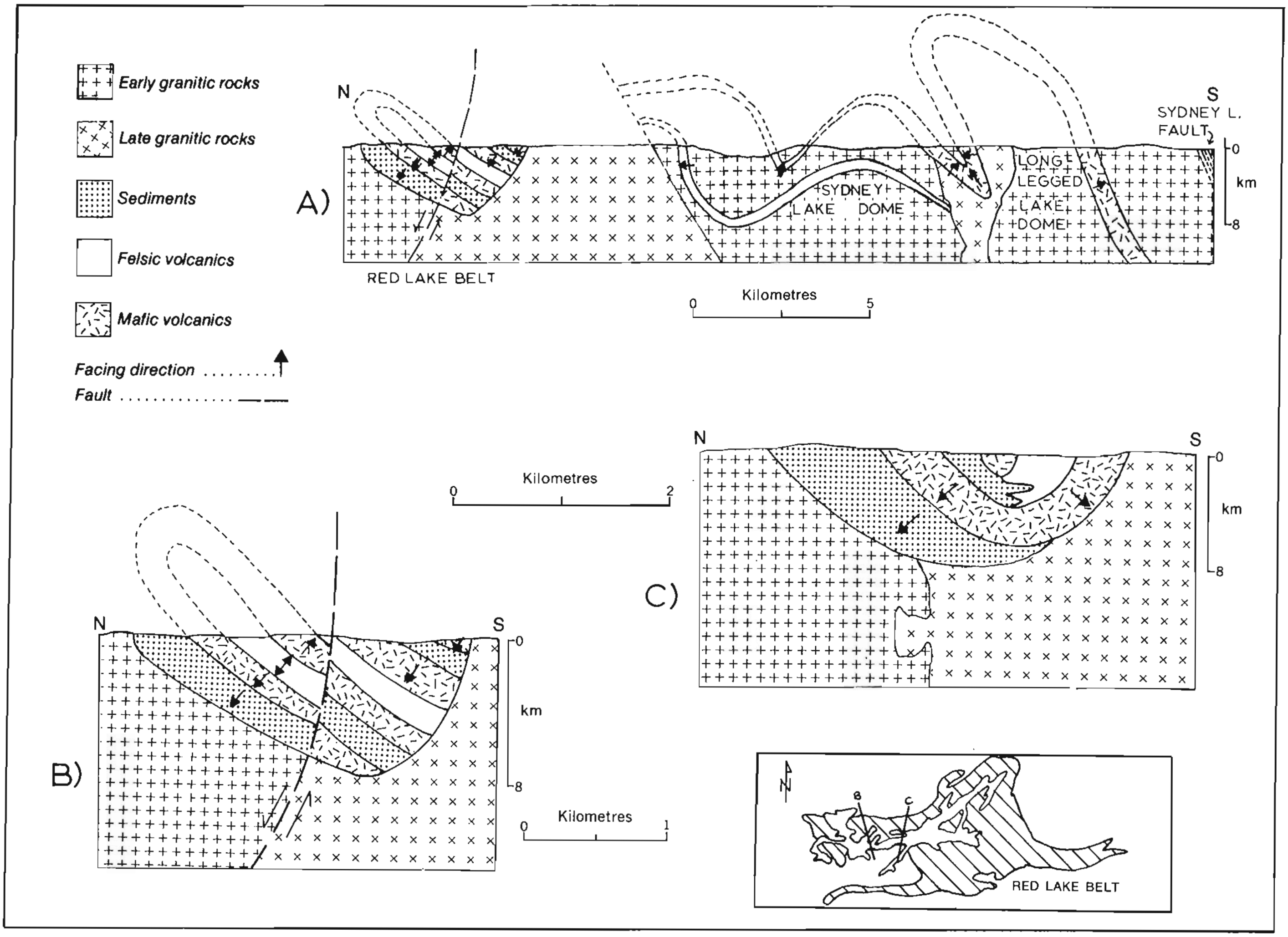

Figure 5. Cross-sections of the Red Lake belt and the Sydney Lake and Rainfall Lake gneiss domes. The volcanic fragments preserved at the margins of the dome younging outward represent the lower portion of the nappe and the Red Lake belt the nose of the nappe.

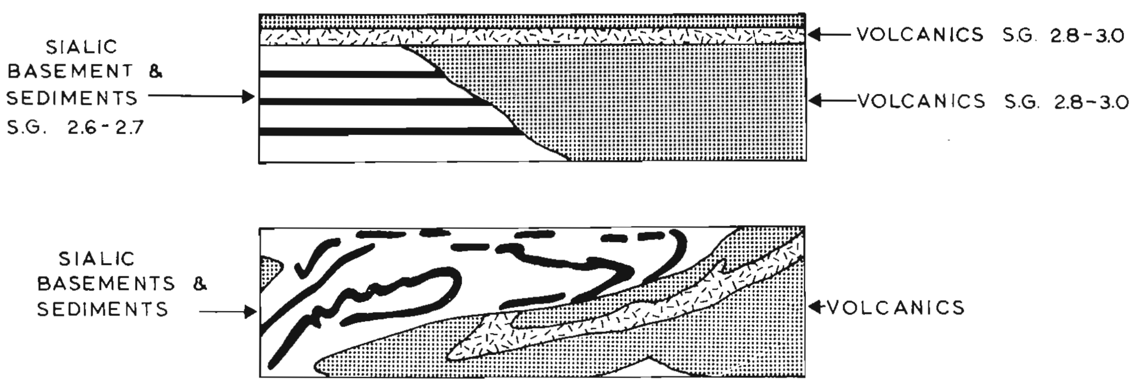

Figure 6. Reconstruction of the English River Subprovince-Uchi Subprovince interface before thrusting.

SUMMARY AND CONCLUSIONS

The map of metamorphic zones is dominated by a pattern of low grade synclinally folded metavolcanic-metasedimentary belts in the Cross Lake subprovince succeeded to the south by the larger, more complex generally low grade metavolcanic-metasedimentary belts of the Uchi subprovince which range in age from 2960 Ma to 2740 Ma (Nunes and Thurston, 1978). The Uchi subprovince is bounded to the south by the Sydney Lake cataclastic zone succeeded southward by the metasedimentary and plutonic terrain of the English River subprovince with both domains containing granulite grade areas not exceeding 3080 Ma in age (Krogh et al., 1976).

The metamorphic zones recognized are as follows: low grade – chlorite-biotite zone; medium grade – staurolite chlorite-biotite zone, sillimanite-muscovite zone; high grade – sillimanite -K feldspar zone, cordierite-almandine zone. Isograds in the area of the Uchi-English River subprovince boundary parallel the Sydney Lake cataclastic zone, with an abrupt rise in grade over a distance of 200 m south of the zone. In the Lake St. Joseph region, metamorphic grade rises southward within the English River Subprovince, ascending to an extensive zone of low pressure granulite facies assemblages which overprint the interdomain boundary. The pattern of metamorphic zonation is explicable in terms of distribution of areas of high heat-flow. Richardson's (1970) model of thermal anticlines within metamorphic belts best fits the zonation pattern.

The east-trending Sydney Lake fault system lies near the transition from the mainly metavolcanic Uchi Subprovince to the mainly metasedimentary English River Subprovince. Branches of the fault curve northward into the metavolcanic rocks of the Uchi Subprovince. The Sydney Lake cataclastic zone and its various branches are proposed as thrust faults, based upon abrupt changes in metamorphic grade associated with the faults, the presence of homoclinal metavolcanic sequences, marginal anticlines in the metavolcanic belts, and hematite along the fault.

The homoclinal, overturned Bee Lake metavolcanic-metasedimentary belt, the anticlinally folded Red Lake metavolcanic-metasedimentary belt, and the mantled gneiss domes south of Red Lake are postulated to be allochthonous fragments derived from northward tectonic transport of recumbent nappes.

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METAMORPHIC HISTORY OF THE ARCHEAN ABITIBI BELT

Wayne T. Jolly[1]

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Abstract

Phase distributions, compositions, and interactions in all major rock types of the Abitibi Belt suggest five periods of metamorphism. The earliest rocks are preserved within retrograded interplutonic amphibolite-orthogneiss in granitic terranes. Sialic sedimentary belts, younger than the gneiss but apparently older than the volcanic rocks of the greenstone belts, were also metamorphosed to amphibolite facies and thus contrast with adjacent volcanic-sedimentary terranes which are of lower metamorphic grade. The sialic sediments are assumed to have been deposited, deformed, metamorphosed, and eroded during the interval from 3600 to 2950 Ma. Barrovian facies series assemblages, identified in sialic metasediments that form the northern border of the Abitibi Belt and in Pontiac schist which occur on the southern boundary of the Belt, indicate high temperatures and relatively high P_T during metamorphism. Eruption of volcanics on the old gneissic complex during the interval 2950 to 2750 Ma was accompanied by downwarp, block faulting, and relatively low pressure burial metamorphism. In addition, lower greenschist metamorphism occurred near subvolcanic magma bodies. A period of erosion of the volcanics and contemporaneous development of synformal sedimentary basins followed. Localized alkaline intrusion and eruption yielded greenschist to amphibolite facies alteration in the sediments and underlying lavas at this time. The metamorphic history of the rocks culminated with the Kenoran Orogeny between 2600 and 2400 Ma ago. This continent-wide disturbance resulted in diapiric intrusions, presumably derived by migmatization of the gneissic basement at depth, into all rocks present in the area and yielded wide contact aureoles in facies series of Abukuma type, incompletely overprinted on previous metamorphic assemblages. During the past 2400 Ma, the area has been thermally stable.

Résumé

La distribution et la composition des phases, et l'interaction entre les phases dans tous les principaux types de roches de la zone d'Abitibi montrent qu'il y a probablement eu 5 périodes de métamorphisme. Les roches les plus anciennes sont préservées dans des terrains granitiques à l'intérieur d'un orthogneiss interplutonique à amphibolite, qui a subi un métamorphisme régressif (ou rétrograde). Les zones sédimentaires sialiques, plus jeunes que le gneiss, mais apparemment plus anciennes que les roches volcaniques des zones de roches vertes, ont aussi été métamorphisées dans le faciès amphibolite, et contrastent ainsi avec les terrains volcano-sédimentaires adjacents moins fortement métamorphisés. On suppose que les sédiments sialiques ont été déposés, déformés, métamorphisés et érodés pendant un intervalle se situant entre 3 600 et 2 950 Ma. Les assemblages des séries de faciès barruvien, identifiés dans les métasédiments sialiques qui forment la lisière nord de la zone d'Abitibi et dans le schiste de Pontiac situé à la limite sud de cette zone, indiquent que les températures ont été élevées et que la pression P_T a été relativement forte pendant le métamorphisme. Les éruptions volcaniques qui se sont produites dans l'ancien complexe gneissique pendant l'intervalle de temps qui se situe entre 2 950 et 2 750 Ma, ont été accompagnées de phénomènes de subsidence, de fracturation en blocs, ainsi que d'un métamorphisme d'enfouissement plutôt faible. De plus, un métamorphisme de la base du faciès des schistes verts s'est manifesté près des corps magmatiques subvolcaniques. Une période d'érosion des terrains volcaniques et la formation simultanée de bassins sédimentaires synformes ont suivi. Ensuite, une phase d'éruptions et intrusions alcalines localisées ont provoqué l'altération des sédiments et des làves sous-jacentes déjà mises en place et leur métamorphisme dans le faciès schistes verts et amphibolite. La culmination du métamorphisme a eu lieu pendant l'orogenèse du Kénoranien, il y a 2 600 à 2 400 Ma. Ce bouleversement de tout le continent a favorisé la mise en place d'intrusions diapiriques, probablement dues à la migmatisation du soubassement gneissique profond, dans tous les terrains de la région; dans la série de faciès du type Abukuma, de vastes auréoles de contacts se sont formées, sans complètement effacer les assemblages métamorphiques antérieurs. Depuis 2 400 Ma cette zone est stable du point de vue thermique.

INTRODUCTION

A two-year study of the Abitibi Belt and the associated intrusives has been completed and a metamorphic map of this belt has been compiled using specimens collected at close intervals from traverses of all access routes (Fig. 1). The results of the project are summarized in this paper.

GEOLOGIC SETTING

Lithologic Subdivision

The rocks of the Abitibi Belt may be divided (Table 1) for convenience into four broad groups:

1) Tonalitic gneiss, displaying abundant evidence of retrogressive metamorphism to lower amphibolite or greenschist grade, is abundant between the volcanic-sedimentary complexes and granitic plutons.

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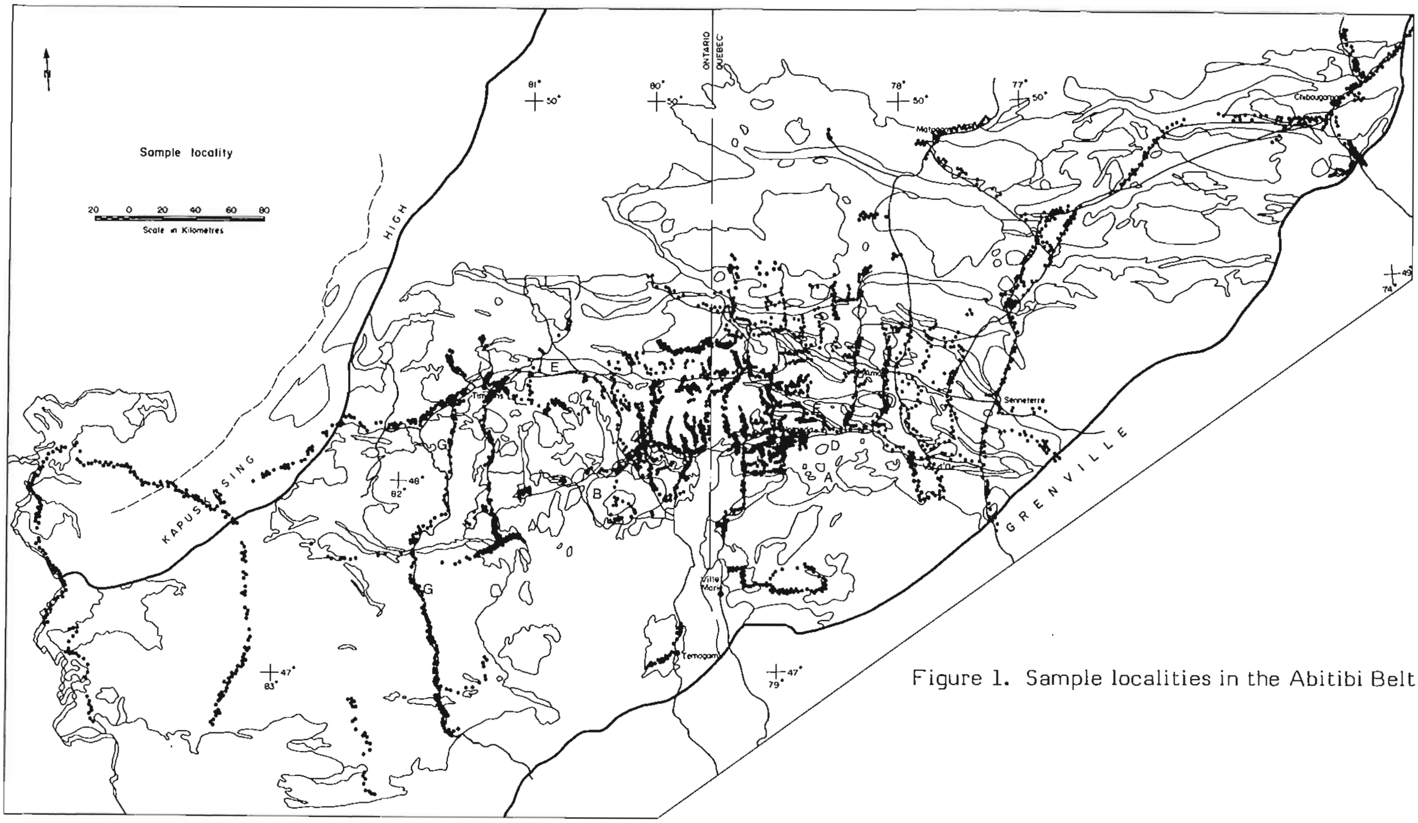


Figure 1. Sample localities in the Abitibi Belt.

2) Eruptive and associated subvolcanic rocks, of great extent and thickness, yielding zircon ages of about 2750 Ma. (Krogh and Davis, 1971), are normally arranged in piles that progressively become more felsic upward (Goodwin, 1974; Jolly, 1975; Gélinas and Brooks, 1974). Commonly they are folded into narrow synclines between large granitic intrusions. The oldest lavas, predominantly mafic and ultramafic flows of Mg-rich komatiitic type, are mixed with moderately Fe-enriched tholeiitic types (Arndt, 1976). They are overlain by a thick sequence of strongly Fe-enriched tholeiites with progressively increasing breccia fractions concentrated at flow tops (Wilson and Morrice, 1977). The upper third of the typical stratigraphic section is composed largely of pyroclastic volcanic cones (Pyke et al., 1968).

3) Sediments, mostly postvolcanic but also pre- or synvolcanic (Dimroth et al., 1973), are commonly present in the volcanic synclinal troughs. Most of the sediment is deep water greywacke; iron formation is common. However, in certain parts of the synclinal troughs, such as in the area near Kirkland Lake, Ontario, abundant textural evidence demonstrates that conglomerate was deposited under subaerial conditions (Hyde and Walker, 1977). Clasts from conglomerates derived from the adjacent volcanic pile carry a prehnite-pumpellyite assemblage which suggests that some of the sedimentary basins were formed on the lava pile after or during the initial episode of folding.

4) Intrusive rocks, broadly circular in plan, are abundant within and on the margins of the lava-sediment pile; they are divided into four broad types:

a) Gabbroic, dioritic, tonalitic, and other intermediate bodies of layered or massive character, are identical in composition to the enclosing lavas (Jolly, 1977a, b). Commonly, the bodies contain anorthositic phases.

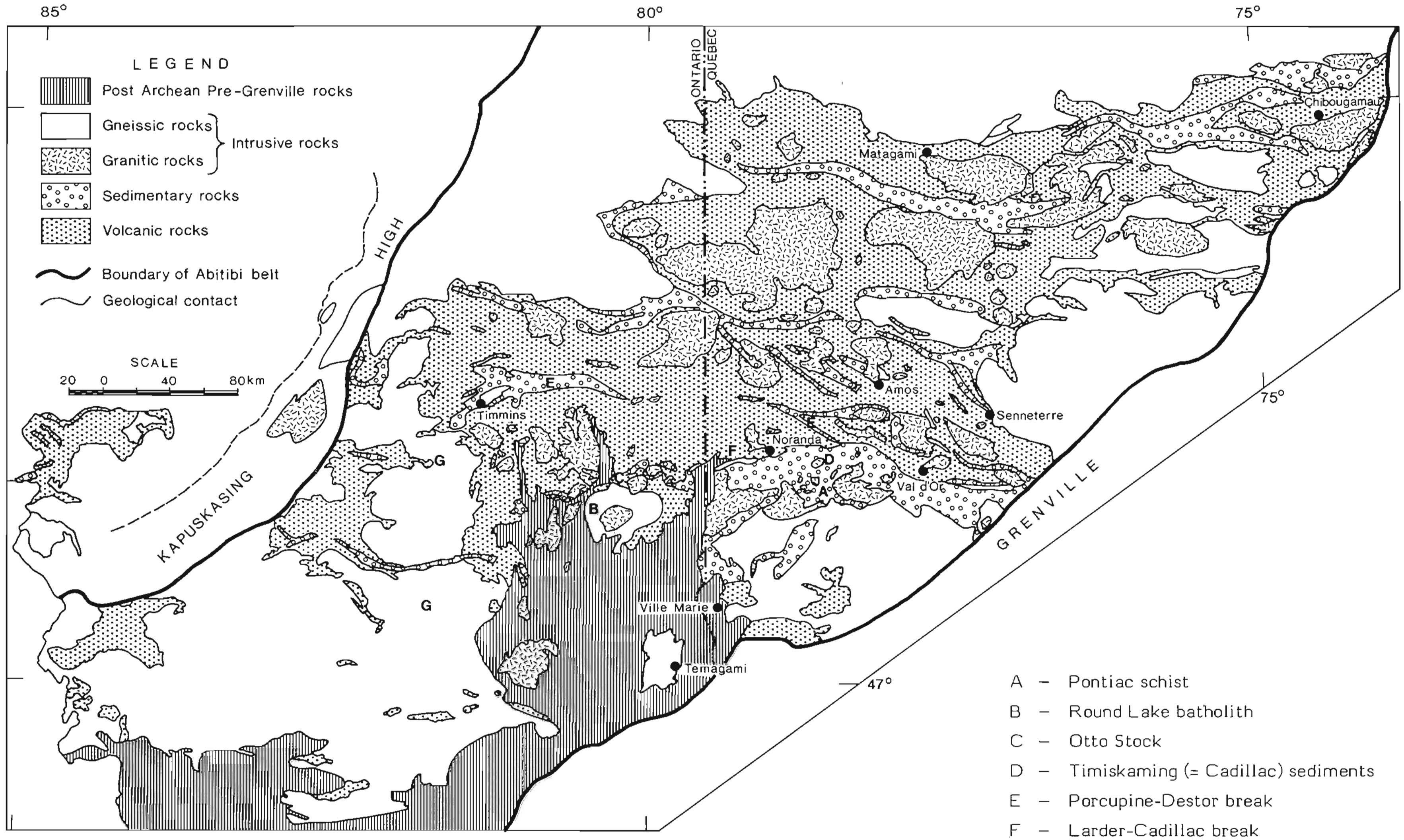
b) Granitic plutons (Fig. 2) of large size are commonly present within the lavas and the surrounding terrane; abundant pegmatite dykes are closely associated with many such plutons.

c) Granitic gneiss plutons (Fig. 2), commonly with igneous interiors, occur in the border zones of the volcanic belts and locally intrude the volcanics. The margins of the plutons are commonly retrograded and contain characteristic greenschist facies assemblages (Ridler, 1972; Jolly, 1974).

d) Late syenitic stocks and accompanying epizonal dykes and sills are widely distributed. Chemical data suggest that these plutons are comagmatic with the alkaline volcanism of the primarily sedimentary Timiskaming Group of the Kirkland Lake area (Cooke and Moorhouse, 1969; Ridler, 1970).

Tectonic History

The volcanic pile and the underlying gneiss were intruded and domed by granitic diapirs of Kenoran age (2400 Ma, see Goodwin and Ridler, 1970). Steeply dipping synclinal troughs between the plutons generally trend east-west. The degree of doming is related to the size of the individual pluton, hence the larger intrusive bodies are surrounded by volcanic rocks from deep within the pile. The stratigraphically highest lavas are preserved only in areas relatively remote from such intrusions. Thicker accumulations of lava such as in the area north of Kirkland Lake and Larder Lake, are not intruded by massive plutons. They behaved as independent tectonic blocks of broadly synformal character during Kenoran deformation.



A – Pontiac schist

B – Round Lake batholith

C – Otto Stock

D – Timiskaming (= Cadillac) sediments

E – Porcupine-Destor break

F – Larder-Cadillac break

G – Gogama traverse

Figure 2. Geological sketch map of the Abitibi Belt simplified from maps of the Quebec Department of Natural Resources and Ontario Division of Mines. The gneissic terranes contain undesignated granitic intrusives equivalent to those shown within the volcanic belts.

The interrelations between the plutons and synclinoria may be readily used to subdivide the Abitibi Belt into tectonic blocks (Fig. 3, cf. Kalliokoski, 1968). Each block is delineated by structural breaks or major synclinal features, commonly containing sediment, which surround uplifted volcanic rocks and the plutonic cores. At least 15 blocks are delineated, each displaying a dominant metamorphic rank which is dependent on the size and type of intruded plutonic core. The tectonic blocks are viewed as centres in which the granitic cores rose to form anticlinal structures, while the denser surrounding lavas sank to form deep, steep-sided troughs.

METAMORPHIC ZONATION OF VOLCANIC ROCKS

The extrusive rocks exposed in the region between Chibougamau, Quebec and Timmins, Ontario, have been divided, for purposes of metamorphic classification, into five zones corresponding to three metamorphic facies (Fig. 4). Each of the five zones is based on microscopic identification of secondary minerals. Most sepcimens, however, may be classified in the field from textural features.

Zone 1

Rocks exhibiting simple diagenesis are not common in the Abitibi Belt. They are found only in sedimentary belts composed largely of greywacke or conglomerate. The best known example is the western part of the Timiskaming Group near Kirkland Lake, Ontario. The matrices, clasts and rock fragments of these sediments commonly carry white mica and clay minerals as the only secondary phases. A few volcanic clasts have been reported to contain detrital epidote, prehnite, and pumpellyite derived from nearby volcanic rocks (Jolly, 1974). These sediments, in part fluvial in origin, unconformably overlie all other Archean rocks in the area except a few plutonic intrusions of syenitic composition. The latter have developed metamorphic aureoles, so that certain parts of the sedimentary deposits contain actinolite or even hornblende.

Subgreenschist facies – Zone 2

The zeolite facies as defined by Coombs et al., (1959) is not present in the Abitibi Belt. Rocks containing prehnite, prehnite plus pumpellyite, or pumpellyite plus epidote, are widespread (Baragar, 1968; Goodwin, 1974; Jenson, 1972; Gélinas and Brooks, 1974; Jolly, 1974) within the volcanic rocks (Fig. 4). Such rocks are most common in the Lake Abitibi region near the Ontario-Quebec boundary, where the greenstone belt is widest. Near intrusive bodies, prehnite-pumpellyite rocks give way to actinolite-epidote-bearing types which originated as a result of low pressure contact metamorphism. Even in areas close to the lobate intrusions at the boundary of the greenstone belt, the metamorphic effects on the volcanics are commonly superficial and suggest little deep-seated activity:

1) The pumpellyite-prehnite-bearing rocks have been overprinted by actinolite-epidote-bearing assemblages in all contact aureoles, suggesting that the former was originally present throughout the lavas before intrusion occurred.

Table 1

Formations and metamorphic events in the Abitibi Belt

Unit/Event	Lithology	Age (Ma)	Metamorphism
Kenoran Orogeny	Granitic plutons	2600-2400	M-5: Contact metamorphism; greenschist and amphibolite facies
Timiskaming Group (=Cadillac Group?)	Greywacke, conglomerate		
	Unconformity		
Stocks	Alkali syenite, syenite		M-4: Contact metamorphism; greenschist and amphibolite facies
	Intrusive Contact		
Abitibi Group	Volcanics, minor volcanogenic sediments	2950-2750	M-3: Burial metamorphism; lower greenschist facies near magma bodies
	Unconformity		
Pontiac Group (=English River and Quetico gneisses?)	Sialic sediments and metasediments	3600-2950?	M-2: Regional metamorphism; intermediate pressure amphibolite facies
Orthogneiss	Tonalite, amphibolite	3900-3600?	M-1: Regional metamorphism; amphibolite facies
Early Archean Crust?		+3900?	

2) Higher grade terranes commonly contain irregular patches, more than 1 km wide, in which pumpellyite is preserved; this is a common features of hornfelses derived at low pressures (see Turner, 1968).

3) Prehnite-pumpellyite-bearing rocks are neither foliated nor penetratively deformed, and therefore retain original textures; plagioclase laths, olivine euhedra, and matrix components are recognizable because the secondary minerals do not cross grain boundaries.

4) Most of the pumpellyite or prehnite-bearing rocks contain 4 to 6 per cent H_2O, suggesting that they were hydrated during alteration. Rock hydration occurred predominantly in porous rocks (amygdaloidal zones, flow breccias, pyroclastic rocks), and along fractures, pillow margins, flow tops, or other channelways. Impermeable rocks (flow interiors) contain few secondary alteration products. Such relations suggest that both fluid and load pressures were low during metamorphism.

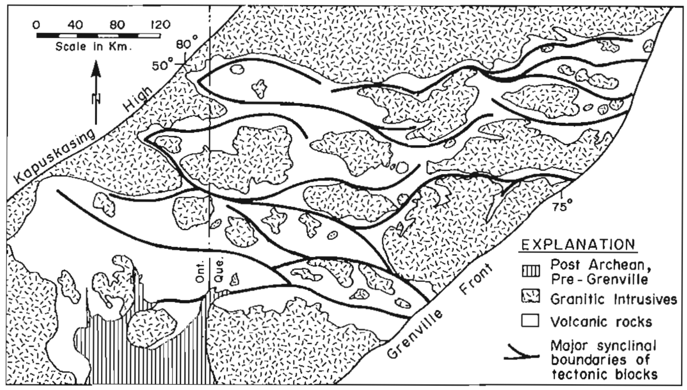


Figure 3. Tectonic subdivision of part of the Abitibi Belt, indicating approximate positions of major synformal features or structural breaks.

Greenschist facies – Zone 3

Rocks of this group contain albite, actinolite, epidote, chlorite, and stilpnomelane as the dominant secondary phases. Actinolite needles are commonly visible in hand specimens, especially where associated with light-coloured veins of calcite, prehnite, or quartz. Most of these rocks display a crude foliation, defined by actinolite growth across primary grains; original textures are blurred and, at higher grade, are obliterated by complete recrystallization. All such metamorphism in the Abitibi Belt is broadly associated with intrusives rather than with depth of burial (see Ayres et al., 1970).

Amphibolite facies – Zones 4 and 5

Amphibolite facies rocks, hornblende-bearing and invariably schistose, were observed only on the margins of intrusive bodies. In this area the composition of the amphiboles is used to separate the lower grade Zone 3 from the amphibolite facies Zone 4. Hence, hornblende and actinolite are distinguished on the basis of pleochroic characters; brownish amphiboles are considered to be aluminous and therefore to be hornblende. Many of the largest intrusions are surrounded by thin zones containing diopsidic pyroxene, garnet, plagioclase, anthophyllite, and less commonly, cummingtonite. In some areas anthophyllite and garnet occur in the amphibolite zone. In general, the widths of the hornblende zones are proportional to the size of the plutons.

Mineralogy of Low Rank Metamorphism

The major characteristic of the widespread low grade metamorphic zone is its patchy development and the tendency of the most complete mineralogical transformations to occur in breccias, amygdaloidal parts of flows, and along fractures. Such features are common in the zeolite and prehnite-pumpellyite facies (Levi, 1969; Jolly, 1970; Jolly and Smith, 1972), where rock alteration is directly related to permeability (Dickinson, 1962). Within the Timiskaming sediments, only the clasts contain prehnite-pumpellyite, suggesting that the clasts were derived from the volcanics following regional metamorphism (Jolly, 1974).

The low grade volcanic rocks are within the prehnite-pumpellyite facies of Coombs (1960), except in isolated localities near intrusive bodies where narrow lower greenschist facies metamorphic aureoles have overprinted the low temperature minerals. No zeolites have been recognized in any of the rocks. Common mineral associations (Fig. 5) have been described elsewhere (Jolly, 1974). Representative electron microprobe data from secondary phases are given in Table 2.

Pumpellyite within any rock is wide ranging in composition and, careful chemical studies must be made to understand its environment of occurrence (Zen, 1974). Generally, pumpellyite occurs in two modes: 1) within amygdules, veins, or intrabreccia cements, usually in association with quartz, prehnite, chlorite, and calcite;in contact with quartz, the mineral is euhedral, acicular to spindly, strongly zoned, and usually contains less Fe than other pumpellyite grains (Fig. 6). 2) With sphene as an extensive groundmass or plagioclase replacement consisting of a fine felt of Fe-rich needles.

Prehnite, which is most abundant in veins and amygdules, or in association with albite as a plagioclase replacement, is also variable in composition. Fe-rich rocks contain radiating clusters of prehnite, which is progressively coloured from brown centres to clear edges; this reflects variations in Fe-content (Fig. 6). Prehnite commonly fills voids embayed by quartz or epidote, especially in rocks near small intrusive bodies.

Chlorite, with low birefringence, and pleochroism ranging from faint green, yellow, and various shades of brown to reddish brown, is present in all rocks as an extensive felty matrix replacement, as pseudomorphs after olivine, and as elongate sheaves lining amygdules and fractures. Sphene is a ubiquitous associate. Most chlorite displays low relief and in many places, anomalous blue birefringence. Some brownish chlorite has first order birefringence and relatively high relief and may consist of interlayered "vermiculite" similar to that described in prehnite-pumpellyite-bearing rocks elsewhere (Jolly, 1970; Zen, 1974). No clay minerals other than vermiculite have been identified.

Epidote and pumpellyite occur together without reaction in relict patches in the southern part of the area. There, rocks from deep within the lava pile are exposed within the broad lower greenschist aureole of the Round Lake Batholith (Fig. 2). This pumpellyite-epidote therefore probably represents a metamorphic grade higher than other pumpellyite-bearing rocks; the epidote (pistacitic) has Al/Al + Fe ratios averaging 0.74 (Fig. 6). The deep lemon yellow pleochroism of these epidotes contrasts with the faint brown, nonpleochroic character of epidote within the closely associated actinolite-bearing assemblages.

Albite, nearly pure and of intermediate to low structural state, replaces most plagioclase phenocrysts and microlites, and commonly forms new growths in veins and amygdules. Plagioclase relicts are preserved in only a few, commonly coarsely crystalline, rock types. Quartz is abundant in all rocks, as are hematite, sphene, and white mica (as traces of sericitic plagioclase replacement).

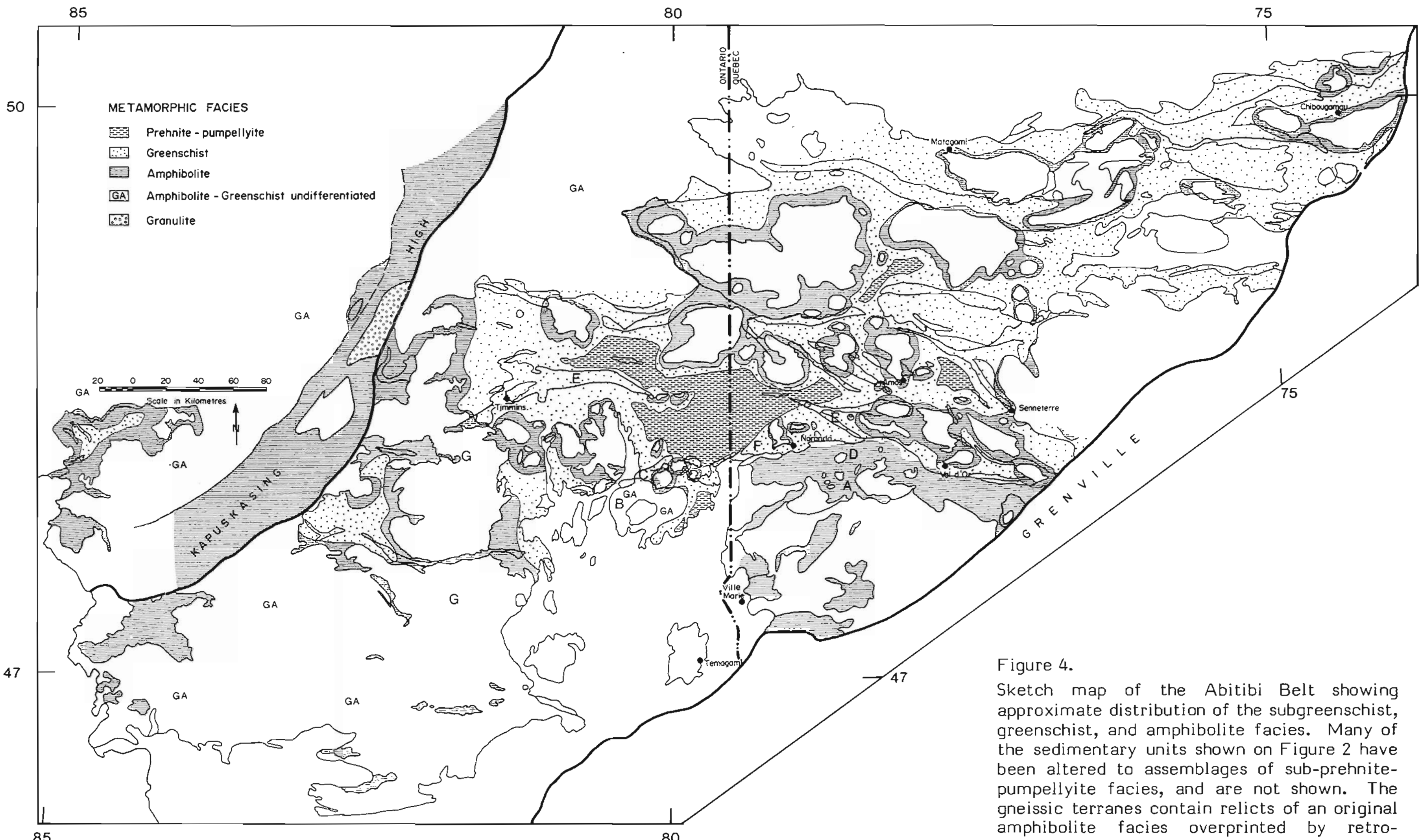


Figure 4.

Sketch map of the Abitibi Belt showing approximate distribution of the subgreenschist, greenschist, and amphibolite facies. Many of the sedimentary units shown on Figure 2 have been altered to assemblages of sub-prehnite-pumpellyite facies, and are not shown. The gneissic terranes contain relicts of an original amphibolite facies overprinted by retrogressive greenschist facies metamorphism caused by postvolcanic granitic intrusion.

Calcite, commonly in intimate contact with Ca-Al silicates, contains pumpellyite and prehnite euhedra in several places, suggesting contemporaneous growth. Calcite also commonly replaces prehnite in the small low grade metamorphic aureoles, implying that a second calcite generation is present.

The relations between mineral and bulk rock chemistry are shown in Figure 6, an Al_2O_3-MgO-FeO_t plot of the major phases. Phases in contact are connected by tie lines. All data points represent averages of eight to twelve points within individual zoned grains from both matrix and void-filling material. The tendency for matrix replacements to contain higher Fe is illustrated. The minerals were all taken from prehnite-pumpellyite-chlorite-bearing rocks of similar metamorphic grade and their compositions are all dependent on bulk rock chemistry. The presence of hematite as an additional phase has shifted bulk rock compositions toward the FeO_t end-member in this projection so that rock data do not fall within the proper three-phase field. Figure 6 also shows relations between the higher grade pumpellyite-epidote-chlorite associations of the relict low grade material near the Round Lake Batholith. The rocks are considerably enriched in iron and, as a result, the chlorites are more Fe-rich than chlorites in lower grade materials of Figure 6. The variation in pumpellyite composition in any single grain is not pronounced; brownish, high-Fe pumpellyite was not observed, perhaps because of the presence of pistacitic epidote.

Mineralogy of Actinolite-Bearing Contact Metamorphic Aureoles

The prehnite-pumpellyite rank regional metamorphism within the lavas has been overprinted during numerous intrusive events (Fig. 4). Most of these lavas contain albite-actinolite-epidote-chlorite assemblages. The aureoles are generally small and associated with stocks and sills, which are also commonly pumpellyitized and of chemical composition similar to the enclosing lavas. In a few of the aureoles rare high grade minerals such as hornblende and plagioclase form thin zones around the plutons. The previously existing prehnite-pumpellyite minerals commonly have been destroyed or deeply embayed by the new aureole minerals (Jolly, 1974; Fig. 4). Biotite is absent from the aureoles except in felsic rocks because of low K_2O content of the lavas; numerous rhyolites in the volcanics near Noranda contain biotite and a biotite zone is also present in the Timiskaming sediments (Cooke and Moorhouse, 1969).

Table 2

Representative partial analyses of major Abitibi volcanic secondary phases (atomic proportions, recalculated volatile free, or structural formulae)

Sample Atomic proportions	220	365
Pumpellyite	1	2
Al	0.41	0.41
Fe	0.13	0.12
Mg	0.06	0.04
Ca	0.39	0.42
mode	vein	matrix
Prehnite	1	
Al	0.46	
Fe	0.04	
Mg	0.00	
Ca	0.50	
mode	matrix	
Chlorite	1	2
Al	0.32	0.35
Fe	0.27	0.41
Mg	0.41	0.22
mode	matrix	matrix
Epidote		1
Al		0.44
Fe		0.16
Mg		0.00
Ca		0.39
mode		matrix
Additional minerals	qtz, cc, sp, ab, se, sp	qtz, ab, he, sp, se

Sample Clinoamphibole, structural formulae	417	412	423
Si	7.76	6.74	7.45
Al	0.32	2.18	0.94
Fe	2.34	2.15	2.06
Mg	2.62	1.94	2.47
Ca	1.97	1.94	2.00
Na	0.19	0.36	0.28
K	0.02	0.06	0.03
Chlorite	3	4	5
Al	0.34	0.34	0.34
Fe	0.42	0.32	0.37
Mg	0.24	0.34	0.29
Epidote	2	3	4
Al	0.44	0.49	0.45
Fe	0.17	0.11	0.15
Mg	0.00	0.00	0.00
Ca	0.39	0.40	0.39
Additional phases	qtz, ab, he, cc, sp, se	qtz, cc, sp, pl, se, mt, he	qtz, cc, sp, ab, he, se

A wide zone of actinolite-bearing rocks near the western and southern margins of the lava pile (Fig. 4), in close proximity to batholithic intrusions, is considered to be within the lower greenschist facies.

Actinolite, generally colourless or faintly pleochroic (bluish green), with birefringence up to first order orange, displays stubby habit with needle-like terminations within rock matrices, and fine grained needles within quartz and calcite void fillings. The mineral is distinguished from pumpellyite by habit, lower relief, bluish tints, higher birefringence, and a tendency to grow across primary grain boundaries as a result of penetrative deformation. Locally, actinolite forms up to 95 per cent of the aureole, and is accompanied by small amounts of epidote, sphene, and chlorite. Plagioclase, pyroxene, and olivine, as well as matrix minerals may be actinolitized. The actinolite crystals contain from 1.5 to 3 per cent Al_2O_3 (Fig. 7) but no trend of increasing Al_2O_3 content toward the Round Lake Batholith was detected.

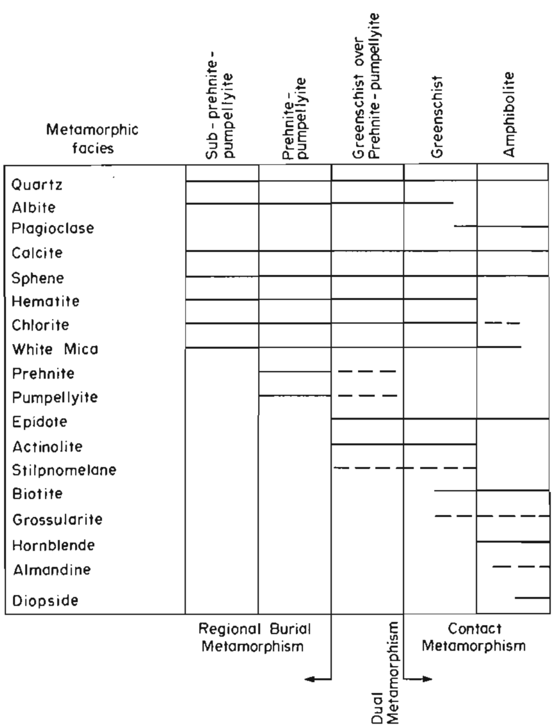


Figure 5. Paragenesis of metamorphic alterations within the volcanic-sedimentary terranes.

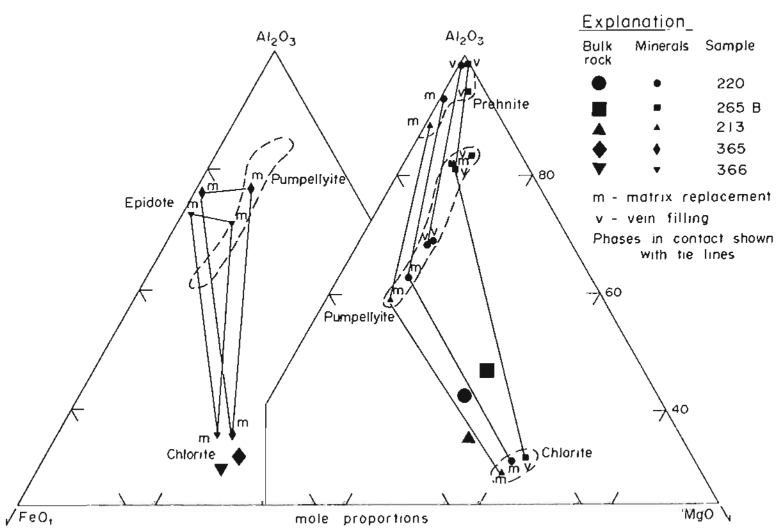


Figure 6. Prehnite-pumpellyite, chlorite, and epidote compositions of the rocks of the Abitibi Belt.

Chlorite in association with actinolite-bearing hornfelses, displays a broad spectrum of colours, from dark green, apple green, yellow green, yellow, yellowish brown, orange to brownish orange. It forms predominantly within matrices, but also replaces both plagioclase and pyroxene crystals; it is commonly absent in highly actinolitized rocks. The Fe/Fe + Mg ratios (Fig. 7) vary widely as a result of differing bulk compositions of the host rocks.

Epidote is widespread and in some rocks forms nearly monomineralic patches as large as 5 by 15 cm along fractures or randomly distributed within massive lavas. Epidote also displays a wide compositional variation, from pistacite having Al/Al + Fe ratios of 0.75 to clinozoisite with Al/Al + Fe ratios of 0.9 (Fig. 7). In the latter, birefringence ranges to 1st order yellow and is nonpleochroic whereas in the former, birefringence approaches 2nd order. Both have exceedingly high relief, granular textures, and commonly display pale lavender anomalous birefringence.

Albite, invariably clouded and peppered by tiny inclusions of white mica, hematite, or various Ca-Al silicates, is nearly free of Ca. No relict feldspar was observed in any lava containing actinolite. Stilpnomelane within the Round Mountain aureole, is generally so fine grained and spindly, that identification is difficult even by microprobe. The mineral, deep brownish orange, is most commonly associated with veins. Andradite garnets, characteristic of veins in the rocks adjacent to the pluton, display slight birefringence (up to 0.003) and strong zoning. It is associated with calcite, actinolite, and epidote, all of which are calcic. Presumably, the garnet formed because of the unusual chemical compositions of the veins. Within the aureoles, quartz, hematite, sphene, white mica, and calcite, are ubiquitous.

The compositions of the diagnostic minerals in the Otto Stock and Round Lake aureoles are illustrated in Figure 7. No compositional trends resulting from metamorphism can be recognized, and the chemistry of the analyzed minerals is attributed to variations in bulk rock chemistry of the original lavas. All minerals analyzed display some zoning within individual grains, but the composition of phases in matrices and void fillings are generally similar. Stilpnomelane coexists with the most Fe-rich chlorite (see also Brown, 1967) and epidote, whereas actinolite is present with Mg-rich chlorite plus epidote. These rocks appear to have more closely approached bulk rock equilibrium during the contact metamorphic phase than those involved in the earlier regional prehnite-pumpellyite metamorphism.

Mineralogy of Hornblende-Bearing Contact Metamorphic Aureoles

Several intrusive stocks throughout the area are surrounded by hornblende-bearing zones at least 1 km wide; smaller intrusive bodies are surrounded by actinolite aureoles only. The Otto Stock of the Kirkland Lake area was chosen for chemical study of such contact effects. This highly alkaline, generally equigranular syenitic stock contains abundant nepheline, especially at the margins, with riebeckite and alkali feldspars. The contact aureole can be divided into several progressive zones which cut across all structures and rocks in the area, with the exception of the nearby Round Lake Batholith.

Zone (1)

The outer metamorphic zone surrounding the Otto Stock is the actinolite zone discussed in the previous section. This zone is classified as belonging to the lower greenschist facies.

Zone (2)

A zone extending 1 to 2.5 km outward from the Otto Stock contains amphiboles more aluminous than those of the Round Lake Batholith, although they are within the actinolite field of Shido and Miyashiro (1958). These amphiboles are distinctly bluish green; brown tones are lacking. Much of this material occurs as cores rimmed by ordinary faint bluish actinolite, probably developed by the later low grade metamorphism of the Round Lake Batholith, an observation supporting data of Ridler (1977). The Otto Stock therefore predates intrusion of the Round Lake Batholith. The pale variety of amphibole is completely eliminated near the stock. Plagioclase (An_5 to An_{50}) and epidote are present in all rocks. The proportion of Al/Al + Fe in epidote ranges from 0.75 to nearly 0.9 and is closely related to bulk rock composition (Fig. 7). Relict minerals are absent but quartz, white mica, calcite, hematite, and anhedral sphene are present as additional secondary phases. A garnet from vein material displays high CaO and FeO_t but low Si/Al ratios. This zone is considered to be within the upper greenschist facies.

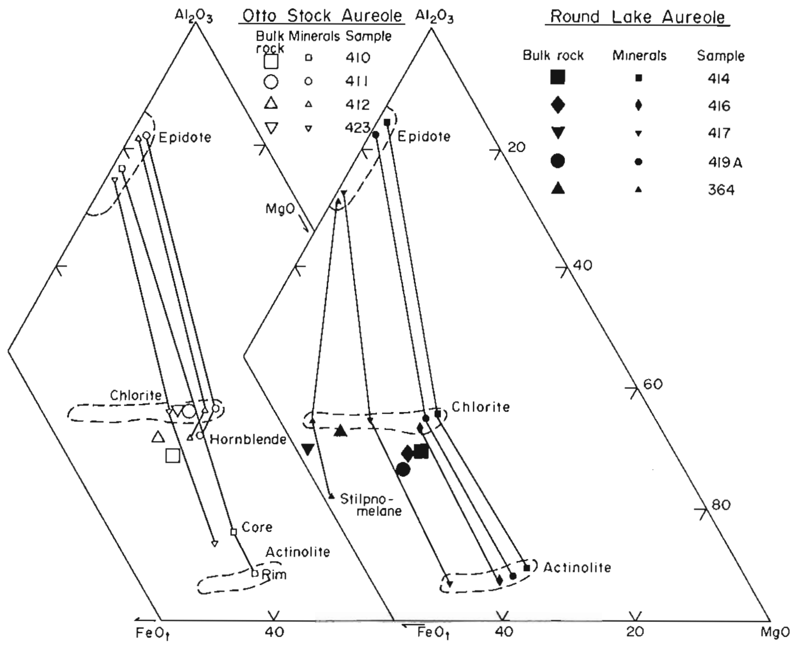


Figure 7. Representative compositions of minerals from the greenschist and amphibolite facies in the Otto Stock and Round Lake aureoles.

Zone (3)

The composition of the amphibole changes progressively within 0.5 to 1.0 km of the pluton. Analyzed amphiboles are in or near the field of common hornblende. One is intermediate between the typical actinolite and hornblende fields of Shido and Miyashiro (1958). This mineral is readily distinguished from the lower grade actinolite by its moderately dark brown to deep greenish blue pleochroism, and displays 1st order red birefringence. The mineral is subhedral to euhedral and large crystals are distinctly bladed in form. Oligoclase to labradorite compositions were determined by universal stage; no albite was observed in rocks carrying brownish hornblende. The presence of abundant clinozoisite, hematite, and andradite, suggests high f_{O_2} during metamorphism, perhaps as a result of hydrothermal fluids introduced into the country rock during intrusion of the volatile-rich alkaline stock. Sphene is generally euhedral and rather large in contrast to its amoeboid character in lower grade rocks. Chlorite is not common, but is present locally. Large amounts of brown biotite, the significance of which will be considered below, is present in many rocks. Most rocks of zone (3) display pronounced schistosity which parallels the margin of the Otto Stock.

Zone (4)

Rock a few tens of metres from the Otto Stock, xenoliths, and a large roof pendant in the cente of the mass, are coarsely recrystallized. Coarse grained hornblende and/or rarely, anthophyllite, are present. Hornblende contains as much as 2 per cent Na_2O and 0.3 per cent K_2O, much more than observed in lower grade amphiboles, a trend described in high grade rocks from many localities (see Seki et al., 1969). Diopside is abundant as large euhedra (up to 1.5 cm long). Biotite is ubiquitous. Quartz, sphene, calcite, magnetite, epidote, plagioclase (An_{25-90}), and K feldspar (in veins) are common. Chemical study of the rocks adjacent to the pluton, compared with similar types elsewhere, suggests considerable K_2O introduction by the alkaline volatile-rich stock.

METAMORPHISM OF THE PONTIAC SCHIST

The Pontiac schist, ranging from quartzose greywacke to pelite and containing interlayered mafic igneous rocks, may be subdivided into four metamorphic zones, which trend east-west parallel to the intrusive contact that forms the southern margin of the metasediments. Zones containing biotite, garnet plus hornblende, staurolite, and kyanite may be distinguished from north to south (Fig. 8). The common appearance of hornblende and garnet, the absence of cordierite, and the appearance of kyanite are typical features of metamorphism under higher pressures in the Barrovian Facies Series (Winkler, 1967). A second lower grade metamorphic event overprinted on much of the unit, is extensively exposed in the south near the contact with granitic rocks. Holubec (1972) and Dimroth et al. (1973) have recorded several igneous plutonic events within the granitic terranes. The earliest must have occurred at great depths, but the second probably occurred following a period of erosion, for muscovite and chloritoid normally replace kyanite, staurolite, and garnet. Hornblende is commonly altered to actinolite or epidote. Compositions of mineral assemblages are illustrated in Figure 9.

METAMORPHISM OF THE PRE-VOLCANIC GNEISS

Most Archean volcanic-sedimentary terranes are underlain by even more ancient gneissic complexes that have been extensively disrupted by diapiric plutons and abundant pegmatitic dykes and other dykes similar to those present within the volcanic terranes. The texture, structure, and compositional uniformity of the gneissic complexes suggest that they may be collectively classified as retrograded amphibolitic facies orthogneiss. Analyses (Table 3) demonstrate that the gneissic-textured rocks are depleted in K_2O, but enriched in CaO, such that there is little overlap between compositional fields of the retrograded gneiss and granitic diapirs, as illustrated on variation diagrams (Fig. 10). The gneiss are grey, medium grained, foliated rocks with abundant fractures, microfolding, and retrogressive metamorphic alterations, and most display evidence of at least two major metamorphic episodes. The first yielded amphibolite grade assemblages. A second episode, the Kenoran Orogeny, yielded lower grade associations in which hornblende, diopside, and biotite are characteristically degraded to chlorite-actinolite-sphene-epidote-quartz assemblages (Jolly, 1974) and calcium-feldspar is commonly replaced by albite and muscovite-epidote intergrowths. In addition, the following features are common: 1) plagioclase crystals are bent, broken, fractured, and subsequently healed by quartz; 2) edges of feldspar megacrysts are commonly rounded as if by cataclastic grinding; 3) biotite is bent, kinked, and shattered, and appears to have been squeezed between adjacent grains; 4) intergrain spaces are annealed by relatively undeformed granular quartz of finer grain size than other minerals; 5) hornblende and biotite have been rotated both on megascopic and microscopic scales to yield indistinct

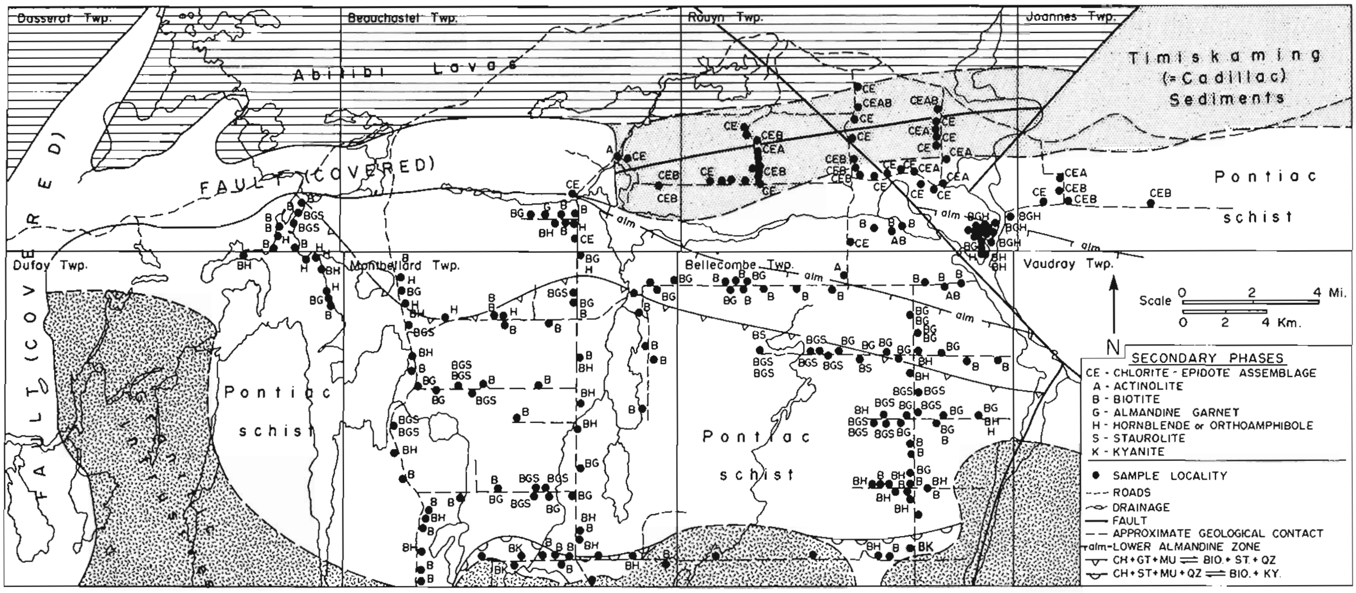


Figure 8. Graphical representation of mineral assemblages and compositions from the Barrovian terranes of the Pontiac schist; data in mole proportions.

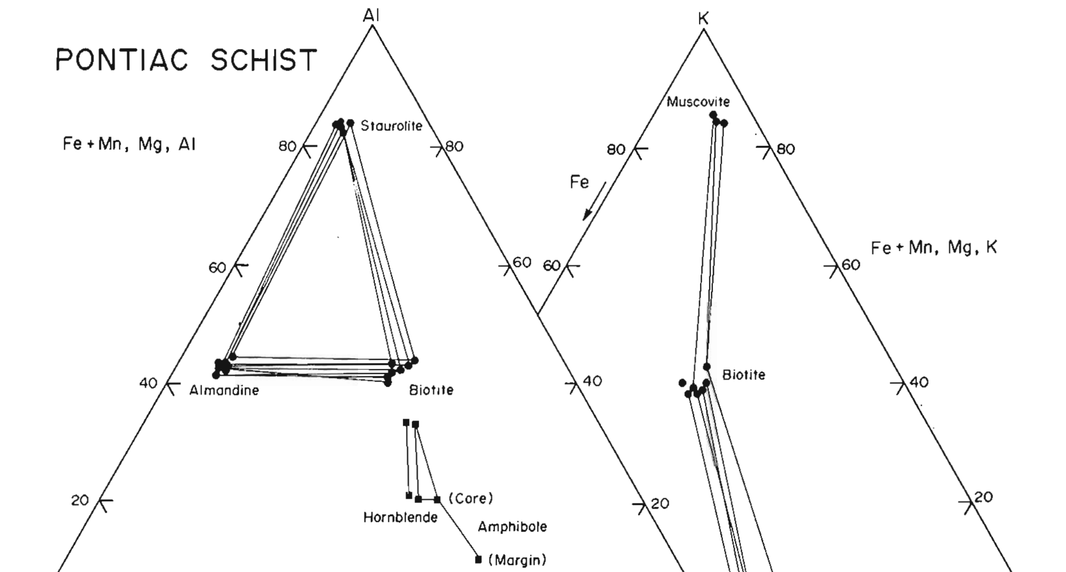


Figure 9. Distribution of metamorphic phases and significant isograds. Most assemblages display overprinted greenschist minerals. Mineral abbreviations (isograds only): chlorite, CH; garnet, GT; muscovite, MU; biotite, BIO; staurolite, ST; quartz, QZ; kyanite, KY.

foliation. Near contacts with true granitic intrusives or in close proximity to pegmatitic and other dykes, the gneiss were commonly recrystallized to hornblende-biotite-(diopside) assemblages that probably closely resemble those of the original metamorphic recrystallization.

The northern border of the Abitibi Belt is bounded by the Quetico Belt, a thick sequence of felsic sedimentary gneiss ranging from lower to upper amphibolite (Goodwin, 1972). These rocks are interpreted to represent a prevolcanic orogenic cycle that included periods of sedimentation, deformation, metamorphism, and erosion. The contact between these gneiss and the volcanic rocks is a fault.

The presence of the Kapuskasing horst, an upthrown block of crustal material (see Kalliokoski, 1968), trending about N45°E parallel with and extending northward from the mid-continent gravity high of Keweenawan age, supplies an opportunity to examine metamorphic rocks typical of those at great depth beneath the exposed gneissic and volcanic sedimentary terranes. These rocks, probably uplifted many kilometres during late Precambrian (Keweenawan or Grenvillian) times, are intruded by granitic plutons. Large areas within the uplifted block carry migmatized amphibolite and granulite facies assemblages (see Ayers et al., 1970) that appear to have been formed by high P_t metamorphism of intermediate to felsic igneous parent rocks (Fig. 10).

Table 3

Representative chemical data from rocks of the granitic terranes; FeO_t = total Fe as FeO. Analyses performed by AAS, W.T. Jolly, analyst

Specimen	2251	2252B	2268	2410
Rock type	Granite	Tonalitic granite	Amphibole gneiss	Kapuskasing Amphibole gneiss
(Traverse)	(S. Gogama)	(S. Gogama)	(S. Gogama)	(Chapleau)
SiO_2	74.2	68.0	62.5	63.2
Al_2O_3	12.4	14.6	15.6	15.5
TiO_2	0.00	0.00	1.08	0.40
MgO	0.51	0.68	1.50	1.53
FeO*	1.48	2.12	6.25	7.01
CaO	0.34	2.60	3.13	5.25
MnO	0.02	0.04	0.08	0.11
Na_2O	4.25	2.67	2.29	0.52
K_2O	4.25	6.13	3.58	3.67
H_2O+	1.25	3.53	1.20	2.36
CO_2	0.13	0.06	0.28	0.10
Total	99.83	100.23	99.49	99.65

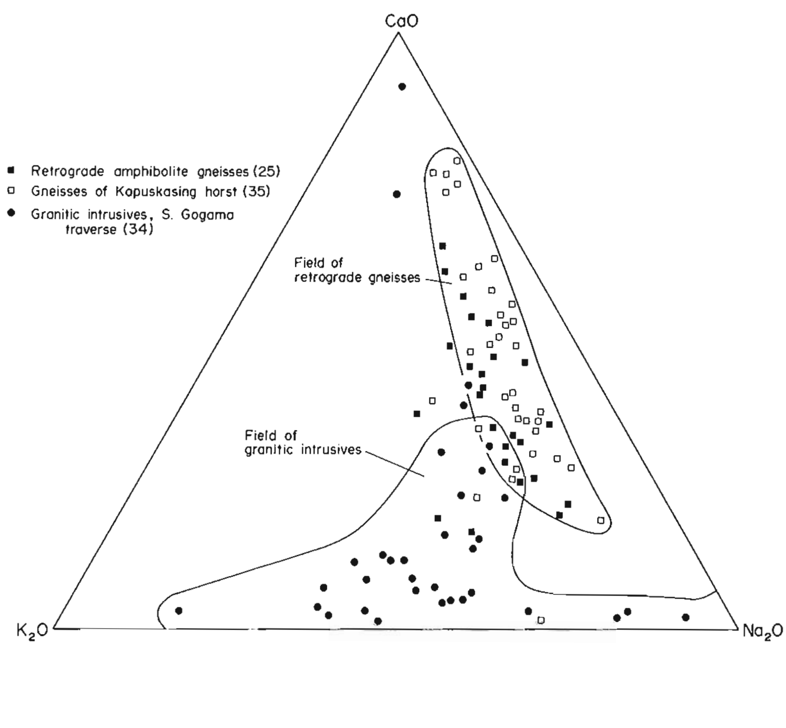


Figure 10.

Na-K-Ca relations of granitic and gneissic rocks of the granitic terranes in the Abitibi Belt.

DISCUSSION

That the Abitibi lavas are generally of deep marine origin has recently been considered in studies of alteration chemistry (Gélinas et al., 1977; Darling and Spitz, 1975); it has been assumed that considerable Na_2O was added to the lavas by exchange with sea water in processes similar to those proposed by Cann (1969), Melson et al. (1968), and Fyfe (1976). Water pressures at depth are great enough to prevent flash expansion of sea water that comes in contact with molten lava, and presumably promotes direct ion exchange between the two (Donnelly, 1966). Some Abitibi lavas contain anomalously high soda, that may represent contamination by the Archean ocean. Such reasoning should not be extended to include more than a small proportion of the lavas, however, for the process of albitization is observed in most low rank metamorphic terranes, including those of demonstrable subaerial origin. The latter type is exemplified by the Keweenawan lavas of the Lake Superior region which display the same metamorphic features as those described here, including similar chemical variations (Jolly and Smith, 1972). Another, probably more significant, difference between these two terranes results from the high water pressures under which the Abitibi lavas were extruded. Sulfur content of the Abitibi lavas is much higher (Naldrett and Goodwin, 1977) than that of the Keweenawan (Cornwall, 1951), because evolution of gas from erupting lavas was suppressed on the ocean floor.

Several aspects of the prehnite-pumpellyite facies metamorphism in the lavas of the Abitibi area, in addition to the absence of the typical high-pressure basaltic replacements, suggest that alteration took place at relatively low pressures. Stilpnomelane has not been reported from the prehnite-pumpellyite facies in strata for which depths of burial were low (Puerto Rico, Jolly, 1970; Tasman Geosyncline, Australia, Smith, 1968; Japan, Seki, 1969), whereas it is present in this facies in areas of great stratigraphic thickness (Japan, Hashimoto, 1970; New Zealand, Coombs et al., 1959; Appalachians, Zen, 1974). Abitibi relations closely resemble those of the above localities. Similarly, the coexistence of actinolite and pumpellyite appears to be restricted to intermediate (Seki, 1969) or high-pressure environments (Hashimoto, 1970; Coombs et al., 1970). Finally, there is a tendency for pumpellyite formed under low P_t to range widely in composition, as does pumpellyite from the Abitibi region, whereas in high pressure terranes the mineral normally has a low Fe-content (Seki, 1960; Coombs et al., 1970) as a result of the presence of other iron-bearing high pressure phases. None of the above relations has been experimentally verified as iron-bearing pumpellyite cannot yet be produced synthetically, but they are observed in rocks similar in composition to those in the Abitibi region, which lends weight to the proposal that the low grade metamorphism concerned here occurred in a volcanic pile of moderate thickness and with a substantial geothermal gradient, as might be expected in rocks of Archean age. The thickness cannot be measured or estimated in the Abitibi area, as a result of: 1) a reliance on attitudes of pillows whose dip may not be determined more accurately than within 30°, 2) the abundance of isoclinal folding and block faulting, together with the patchy exposure available, 3) the interfingering nature of the volcanic rocks whose original surface probably displayed considerable relief near vents that extruded more viscous felsic lavas. Therefore, an estimate of possible thicknesses made from comparison of the metamorphism in this area with known examples, is of interest. Seki (1969, Japan), Surdam (1968, Vancouver Island), and Jolly (1970, Puerto Rico), all observed low pressure sequences with assemblages similar to those of the Abitibi area; strata measured were less than 3 km (1 kb)* thick. Similar relations were observed in the Keweenawan lavas, where depth of the prehnite-pumpellyite facies is about 7 km (2 kb). Conversely, the high pressure features discussed above were observed in South Island, New Zealand (Coombs, 1953) where the measured stratigraphic section was 35 000 feet or about 12 km (3 kb); the jadeite-pumpellyite-epidote-lawsonite-bearing Franciscan terranes (Coleman and Lee, 1963) are considered (see reviews of Ernst, 1971; Turner, 1968) to have originated above the experimentally determined aragonite-calcite inversion pressures or about 30 km (8 kb). The latter pressure cannot apply to Archean terranes, because in these such minerals are absent. It appears from these comparisons that the thickness of the Abitibi lava pile above any of the exposed lavas at the time of prehnite-pumpellyite metamorphism was less than that in New Zealand, or less than about 12 km, a figure that is at variance with those estimated by other methods. Jenson (1976) measured up to 60 000 feet (20 km) of section in the region between the Porcupine-Destor and Larder Lake-Cadillac breaks (Fig. 2) and derived a total of 100 000 feet (31 km) when the lavas north of the Porcupine-Destor break were included. Baragar (1968) measured 45 000 feet (15 km) in the Duparquet area south of the Porcupine-Destor break. Ridler (1970) and Goodwin and Ridler (1970) estimated that at least 40 000 feet (14 km) of lavas are present in the region between the two structural breaks, and that they have been greatly thickened in synformal troughs. Most geophysical models treat the lava piles as bottomless. As the metamorphic characteristics do not reflect burial to such depths during alteration, it is concluded that apparent thicknesses reflect thickening by structural features or topographic irregularities produced during and after volcanism. In support of such reasoning, one may cite the presence of lower grade prehnite-pumpellyite facies on both sides of the Porcupine-Destor break, despite the omission of considerable thickness of lava (Arndt, 1976), which suggests that the younger lavas were never present north of the fault. It is likely that the break represents a hinge line south of which the lavas underwent considerable subsidence during eruption of the upper part of the Abitibi pile. If so, the history of the Porcupine-Destor break is long and complex, as the structure was apparently renewed during intrusion of the Kenoran plutonic diapirs.

The greenschist facies in the Abitibi Belt is also of relatively low pressure origin. It is known most completely in the vicinity of the Round Lake Batholith (Fig. 2), where it constitutes a major proportion of the area shared by thin zones of hornblende and pyroxene hornfels and numerous patches of an apparently earlier epidote-pumpellyite secondary mineralogy. Pumpellyite has not been observed in contact with actinolite but was apparently replaced by it during the generation of the latest metamorphic suite, which includes stilpnomelane and biotite (in rocks with sufficient potash) in addition to albite, chlorite, epidote, and actinolite.

One exception to the otherwise uniformly low pressure metamorphism in the Abitibi area is the sedimentary Pontiac schist in the southeastern part of the map area (Fig. 2). These schist, commonly composed of 90 per cent quartz and feldspar, were originally felsic greywacke, locally conglomeratic. Pelitic interlayers display biotite, hornblende, garnet, staurolite, and kyanite zones similar to the Barrovian-type almandine-amphibolite facies of Winkler (1967). These sediments, almost certainly derived by erosion of sialic, richly quartzose basement, have long been suspected as being the oldest stratified unit exposed (Wilson, 1962). In contrast, the Cadillac (=Timiskaming) sediments are intermediate volcanic greywacke, rarely containing more than 30 per cent quartz. Wilson interpreted this unit as unconformably overlying both the lavas and the Pontiac schist, but the contact relations are debatable (Goulet, 1976). It is clear, however, that the metamorphism of the Pontiac schist was of a completely different order, for the assemblages are indicative of relatively high pressures. The rocks must have undergone extreme uplift, relative to the lava pile. The

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*1 kb = 1 x 10^5 k Pa

Larder Lake-Cadillac break, which consists of a roughly east-west zone of severe synclinal folding, appears to be the northern boundary of this uplifted block. The Timiskaming sediments along the zone are cut by many subparallel south-dipping high angle reverse faults; breccias and cataclastic deformation are commonly associated with these faults. The western limit of the high pressure block, buried beneath later Precambrian tillites, trends almost due north along the Ontario-Quebec boundary (Fig. 2). West of this boundary, lavas of the low pressure metamorphic grades previously discussed are present. The separation probably marks a fault along which the eastern side has been uplifted many kilometres. This fault crosses the Larder-Cadillac break, intersects the lavas, and curves eastward. In the lava pile it separates rocks bearing abundant prehnite-pumpellyite assemblages from the low greenschist assemblages on the southeast.

SUMMARY OF METAMORPHIC EVENTS

The metamorphic history of the Abitibi Belt commenced with the deformation of a relatively thick and continuous Archean sialic crust. The presence of this crust prior to the eruption of the volcanic rocks is supported by: (1) unconformities are present below the lavas (Baragar and McGlynn, 1976); (2) high grade sedimentary gneiss are present at low grade greenstone boundaries; (3) early metamorphosed sedimentary rocks, such as the Pontiac schist, display relatively high P_t (Barrovian) metamorphic assemblages, while the younger lavas and sediments carry low P_t (Abukuma) types; (4) gneiss are abundant in synformal structures throughout the Archean granitic terranes separating the volcanic units; (5) gneiss commonly mantle true igneous plutonic cores of diapiric intrusions within the volcanic areas; (6) the uplifted rocks of the Kapuskasing horst, representative of the material at depth in the region, contain abundant migmatized amphibolite and granulite facies assemblages, suggesting that the sialic component of the crust in the Superior Province of the Shield is extensive both laterally and vertically. The geologic history (Table 1) of the gneiss has been blurred or largely destroyed. Some understanding of these earliest events has been achieved in other parts of the Shield (Bridgwater et al., 1976; Goldich et al., 1970), where radiometric techniques applied to areas of outcrop scale reveal complex sequences of intrusion and metamorphism with ages ranging from 3900 to 3600 Ma. Most rocks of these ages display evidence of thin-skinned thrusting which is presumed to have thickened continental crust to several times its original thickness. These deformations were accompanied by recrystallization of the intrusive complex to gneissic rocks of the upper and lower amphibolite facies. The abundance of relict pegmatite features in the gneiss suggests they may have undergone migmatization at this time; migmatites are absent from later Kenoran alteration products of the Abitibi Belt except for those in the Kapuskasing horst. During the interval between 3600 and 2950 Ma, several ill-defined events of continental proportions occurred. The first of these involved uplift and erosion of the early gneissic complex. Deposited on this gneissic complex was a later sedimentary pile, which eventually produced the Quetico and English River gneiss belts, both of which trend about N80°W across the Superior Province. The high grade metamorphism of the sediments, and the independence of the metamorphism in the lava pile, suggest that these deposits were deformed, metamorphosed, uplifted, and eroded during a 650 Ma period (Table 1) prior to extrusion of the Abitibi lavas. The original extent of the sediments is not clear, as the two belts may represent simple downfolded remnants. It appears likely, in view of stratigraphic distributions and observed Barrovian metamorphic assemblages, that the Pontiac schist (Fig. 2) also represent sediments derived, deformed, metamorphosed, and eroded during this interval (see also Ambrose, 1941; Wilson, 1962; Baragar and McGlynn, 1976).

Submarine eruption of the Archean lavas onto the eroded amphibolite grade gneissic complex took place (Table 1) between 2950 and 2750 Ma (Krogh and Davis, 1971). The lavas, which may have covered large parts of the crust, were downwarped and block-faulted during extrusion. At the same time the lava pile was metamorphosed to prehnite-pumpellyite grade, probably by simple increase in geothermal gradient through burial. Subsequent erosion of the volcanics supplied pumpellyite-bearing clasts to depositional basins developed on the lavas (Jolly, 1974; Hyde and Walker, 1977). The metamorphic and tectonic history was completed by a final cataclysmic continental event, the Kenoran Orogeny, which took place between 2600 and 2400 Ma (Goodwin and Ridler, 1970). This period of north-south compression and crustal convection resulted in gravitational re-equilibration of the crustal block; the gneissic basement, deformed by the denser overlying volcanic rocks, rose to form diapiric plutons with igneous cores of coalesced migmatitic liquids, while the volcanics subsided into the intervening troughs. Near intrusive contacts greenschist and amphibolite facies aureoles were produced in both the greenstone belts and the early gneissic basement.

ACKNOWLEDGMENTS

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NOTES ON METAMORPHISM IN NEW QUEBEC

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Herd, R.K., Notes on metamorphism in New Quebec; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 79-83, 1978.

Abstract

New Quebec is composed chiefly of an Archean craton, bordered by Proterozoic supracrustal sequences, locally overlain by Proterozoic outliers. Most Archean and some Proterozoic rocks have been metamorphosed. The craton was metamorphosed at least twice during the Archean, first to granulite facies and later to amphibolite facies. In the southeast part of the region, a cordierite-garnet granulite grade terrane is preserved; in the northwest, granulite has been overprinted regionally by amphibolite facies metamorphism. Amphibolite grade associations predominate in the Archean rocks; whether most of them derive from the first or second metamorphism is uncertain. Mineral assemblages in the amphibolite facies generally suggest low pressure metamorphism except locally along the margins of the craton, and surrounding the cordierite-garnet granulite area where pressures may have been higher.

Secondary minerals such as chlorite, prehnite and zeolites which imply late subgreenschist to greenschist metamorphism of unknown age, are widespread in Archean rocks and in Proterozoic(?) diabase and gabbro. Greenschist facies metamorphism overprinting amphibolite facies rocks north of the Grenville-Superior Province boundary may be Grenvillian in age.

Proterozoic strata adjacent to the craton are of subgreenschist to greenschist grade, or are unmetamorphosed. One small Sakami Formation outlier contains an andalusite-muscovite-quartz association.

Phanerozoic metamorphism is expressed in breccia and maskelynite around Clearwater Lake, and by sanidinite facies spurrite in Ordovician limestone-marble xenoliths west of Clearwater Lake.

Résumé

Le Nouveau-Québec se compose principalement d'un craton archéen, bordé par des successions supracrustales protérozoïques, localement recouvertes de lambeaux protérozoïques. La plupart des roches archéennes, et certaines d'âge protérozoïque ont été métamorphisées. Le craton a été métamorphisé au moins deux fois pendant l'Archéen, une fois dans le faciès granulite, ensuite dans le faciès amphibolite. Dans le sud-est de la région, subsiste un terrain appartenant au faciès granulite à grenat et cordiérite; au nord-ouest, au métamorphisme du faciès amphibolite s'est superposé régionalement le faciès granulite. Les séries du faciès amphibolite dominent dans les roches archéennes, mais il est difficile de dire si la plupart ont été engendrées pendant le premier ou le second épisode de métamorphisme. Les assemblages minéraux du faciès amphibolite indiquent généralement un métamorphisme de basse pression, excepté localement, le long des marges du craton et autour du secteur des granulites à cordiérite et grenat, où les pressions ont peut-être été plus élevées.

Les minéraux secondaires comme la chlorite, la prehnite et les zéolites, qui indiquent un métamorphisme d'âge inconnu, se situant entre la partie tardive du faciès sous-schistes verts et le faciès schistes verts, sont très répandus dans les roches archéennes et les diabases et gabbros du Protérozoïque. Les roches du faciès amphibolite ont été reprises dans le faciès schistes verts, peut-être d'âge grenvillien, au nord de la limite des provinces de Grenville et du lac Supérieur.

Les strates protérozoïques adjacentes au craton ont été métamorphisées du faciès sous-schistes verts au faciès schistes verts, ou n'ont pas été métamorphisées. Un lambeau peu étendu de la formation de Sakami contient l'association minérale andalousite-muscovite-quartz.

Le métamorphisme du Phanérozoïque a créé des brèches et maskelynites aux environs du lac Clearwater, et des xénolithes de marbre ordovicien à l'ouest du lac Clearwater contiennent de la spurrite, dans une zone du faciès sanidinite.

INTRODUCTION

The distribution of metamorphic facies in most of New Quebec was outlined using minerals and mineral associations identified by petrographic examination of approximately 1200 thin sections, supplemented by published and unpublished data. This information was compiled mainly for the central Archean area (Fig. 1), and also for adjacent areas of the Cape Smith Fold Belt, Nastapoka Arc, Otish Mountains, Grenville Province, and Labrador Trough. Geological Survey of Canada collections and field notes re-examined were those on which the reports of Kretz (1960), Eade (1966), and Stevenson (1968) were based. Data and thin sections were also lent by J.H. Remick, T. Clark, and A. Franconi of the Ministère des Richesses Naturelles du Québec, as were data gathered for the Ministère by J. Wallach and J. Bourne. Published Federal-Provincial aeromagnetic maps at 1:250 000 scale and recent geological compilations (Dubé et al., 1976) were used to delineate metamorphic boundaries in the southern portion of the region.

The present metamorphic map is based on reconnaissance geological mapping and therefore only indicates broad areas of facies distribution; relationships amongst the facies and isograds cannot be specified.

[1] Geological Survey of Canada

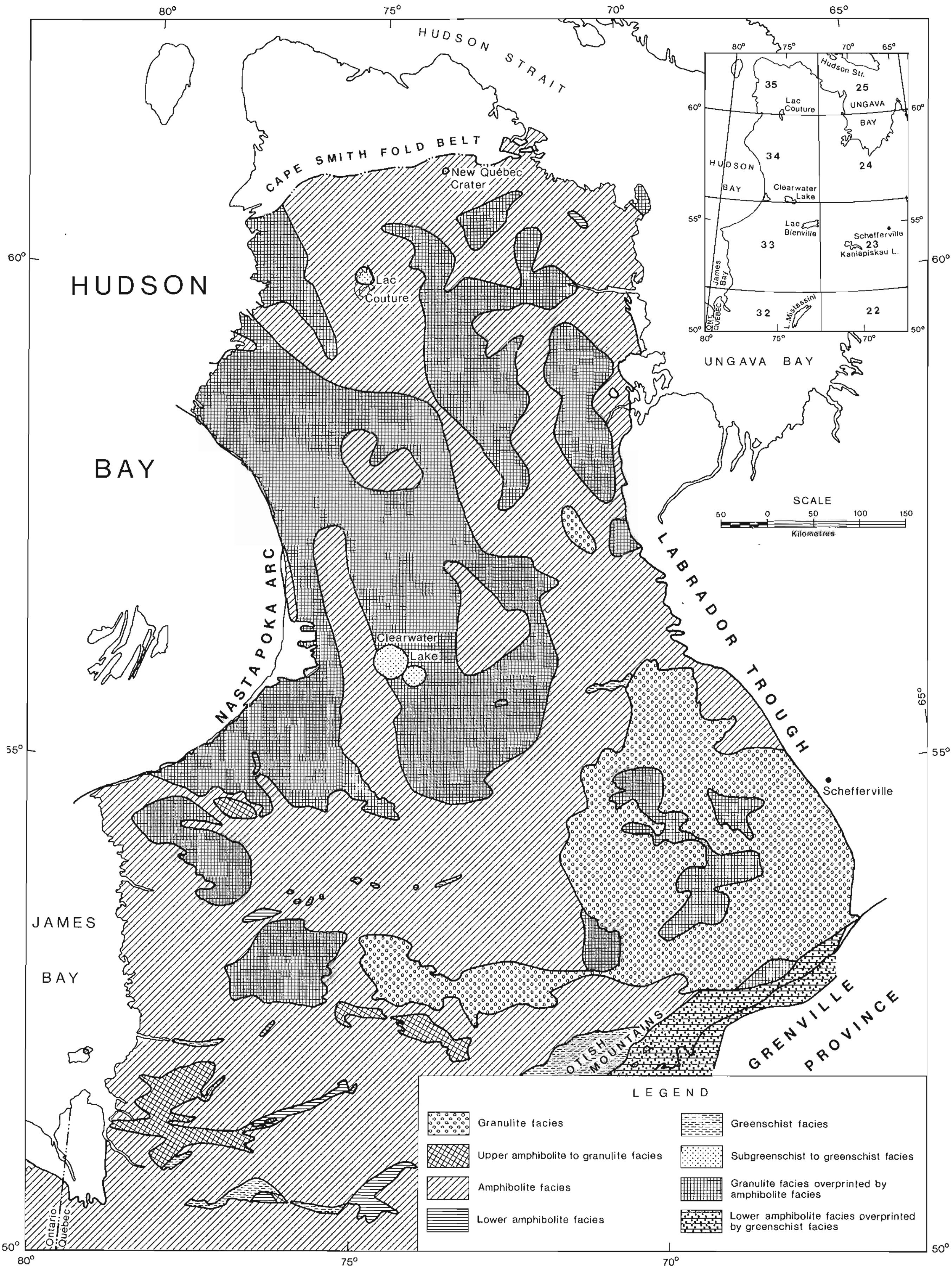

Figure 1. Distribution of metamorphic facies.

Table 1

Selected Archean Metamorphic Mineral Associations, New Quebec

Association																	Grade	Reference	NTS
Gneiss, Schist																			
OX							BF					QU	PA	PE			5	266-4	23C/NW
OX	GA						BF					QU	PC	PE			5	268-25	23C/NE
(OX)CX	GA	HB					BF					QU	PC	PF			4/5	270-16	23C/SE
OX	GA		CD				BF					QU	PA	PE			5	272-3	23D/NE
OX CX							BF					QU	PC	PE			5	289-13	23N/NW
OX	GA		(CD)				BF			CL	(EP)	QU	PA				2/5	313-13	33K/NW
CX	GA		(CD)				BF			CL		QU	PC			SP	2/5	417-35	34L/NE
	GA		(CD)	SL			BF		(MU)			QU	PC	PF			45	280-21	23F/NW
	GA						BF					QU	PA	PE			45	CP-279-61	23J/NW
	GA		CD				BF	(PR)				QU	PA	PE			45	281-34	23K/NE
	GA				AD		BF			CL			PC	PE		SP	3/45	411-39	34G/NW
	GA		CD	SL			BF					QU	PA	PE			45	418-51	34J/NW
			CD				BF			(CL)		QU	(PC)				35	421-16	34N/NW
			CD	SL			BB			CL		QU	PC				2/35	423-55	34O/NW
	GA		CD	SL			BF					QU	PC				35	430-70	35A/SW
	GA		CD				BF					QU	PC				35	433-27	35C/SE
							BB		MU		(EP)	QU	PC	PF			34	266-28	23C/SW
		HG					BF	(PR)	(MU)		(EP)	QU	PC			(CB)	2/34	276-1	23E/SW
		HG					BB			CL	EP	QU	PC	PE			2/34	406-2	24D/SW
	GA						BB					QU	PC				34	426-24	25C/NW
	GA						BB			CL	EP	QU	PC	(PF)			2/34	291-19	33A/NW
			CD	SL	AD		BF					QU	PC				34	KP-64-75	33D/SE
	GA		CG	CD			BB		(MU)			QU	PC				34	R-103-75	33D/NE
	GA						BB	(PR)		CL	EP	QU	PC				2/34	305-33	33H/SE
	GA		CD		AD		BB			(CL)		QU	PC				34	198-1	34P/NW
		HG					BG				EP	QU	PC				3	284-5	23L/SW
	GA		CD	SL		ST	BF			CL		QU	PC				2/3	AF-640-76	33C/SW
				SL	AD		BF		MU			QU	PC				3	JR-119-75	33C/SW
			(CD)		AD	ST	BF	(PR)	MU	CL		QU					2/2-3	300-12	33N/SE
	GA		CD	SL	AD		BF		MU	CL		QU	PC				3	411-59	34G/NW
				SL					MU			QU				TM	3	195-3	34N/SE
	GA		CD	SL		ST	BB			CL		QU	PC				2/3	423-59	34O/SW
	GA			SL		ST	BB			(CL)		QU	PC				3	424-12	34O/SE
				SL	AD	ST	BB		MU			QU	PC				3	424-13	34O/SE
	GA		CD	SL	AD	ST	BB		MU			QU	PC				3	424-17	34O/SE
				SL			BB		MU	(CL)	(EP)	QU					3	424-23	34O/SE
	GA		(CD)	SL			BB		MU	(CL)		QU	PC				3	429-38	35A/SE
			CD	SL		ST	BB			(CL)		QU	PC				3	430-62	35A/SW
										CL	EP	QU	PC	(PF)			2	293-19	33B/SW
Iron formation, Chloritoid Schist																			
OX(CX)			GN				BB					QU			MT		4-5	268-5-2	23C/SE
		HB	CG									QU			MT		34	415-28	24L/NE
			CG							(CL)		QU			MT		34	313-18	33K/NW
		HB	GN									QU			MT	SM	2-3	297-35	33F/SW
									MU	CL		QU				CT	2	304-12	33G/NW
Mafic rocks, amphibolite, metagabbro																			
OX CX		HB					BF						PC				34/5	269-13	23C/NE
OX CX		HB											PC				34/5	408-21	25D/SW
	GA	HG	CG							CL		QU	(PC)			TM	34	274-19	23D/NW
		HG					BB						PC				34	276-10	23E/NW
	GA	HG						(PR)		CL	(EP)		PC			(CB)	2/34	276-15	23E/SW
CX		HG						(PR)				QU	PC			(CB)	2/34	276-2	23E/SW
		HG						(PR)		(CL)	(EP)	QU	PC			(PL)	1/34	279-1	23F/NW
		HG					BB	(PR)					PC				34	286-29	23M/SW
	GA	HG	CG						MU	CL		QU	PC			TM	2/34	286-34	23M/SW
CX	GA	HG											PC				34	406-29	24D/NW
	GA	HG					BB			CL		QU	PC				2/34	AF-676-76	33C/SE

Notes: Associations: Selected to show the range, not the extent or frequency of occurrence; mineral associations are arranged, within each group of rock types, by decreasing grade and by NTS.

Abbreviations: OX:orthopyroxene; CD:clinopyroxene; GA:garnet; HB:brown hornblende; HG:green, blue-green hornblende; CG:cummingtonite; GN:grunerite; CD:cordierite; SL:sillimanite; AD:andalusite; ST:staurolite; BF:fox-red biotite; BB:brown biotite; BG:green biotite; PR:prehnite; MU:muscovite; CL:chlorite; EP:epidote group; QU:quartz; PA:antiperthite; PC:plagioclase; PE:perthite; PF:potash feldspar; MT:magnetite; SP:spinel; CB:carbonate; TM:tourmaline; SM:stilpnomelane; CT:chloritoid; PL:pumpellyite; parentheses denote relatively minor constituent.

Grade: Determined from given mineral associations. 1:Subgreenschist; 2:Greenschist; 3:Lower Amphibolite; 4:Upper Amphibolite; 5:Granulite. Transitional facies indicated by dash: e.g. 2-3, greenschist to lower amphibolite; undivided facies indicated by bar: e.g. $\overline{34}$, undivided lower to upper amphibolite; overprinted facies indicated by slash: e.g. 2/5, greenschist overprinting granulite. Map units in Figure 1 include all grade categories shown in table.

Reference: Specimen numbers (Geological Survey of Canada, and Ministère des Richesses Naturelles, Québec).

Petrography by R.K. Herd and N.A.C. Rey, 1977.

GENERAL GEOLOGY

Detailed discussions of the geology of parts of the area have been published by Eade (1966), Stevenson (1968), Taylor (1975), Taylor and Skinner (1964), and Clarke (1968). The central area of New Quebec is occupied by an Archean craton composed of rocks that have yielded Kenoran and older ages. Peripheral to this craton are Proterozoic supracrustal sequences which may have counterparts in isolated outliers on the craton. Younger (K-Ar) ages have been obtained from basement rocks near the Labrador Trough and the Cape Smith-Wakeham Bay Belt (Wanless, 1970). Recognizable Archean supracrustal belts are found mainly in the southern portion of New Quebec (Dubé et al., 1976), but descriptions by Stevenson (1968) of the Leaf River area show that supracrustal remnants are ubiquitous within gneiss of the 'Ungava craton' (Dimroth et al., 1970). Unmetamorphosed posttectonic granite intrusions are relatively scarce. Diabase dykes and minor intrusions are numerous. The only Phanerozoic rocks are lava, limestone, and contact-metamorphosed inclusions, at west Clearwater Lake (Bostock, 1969).

METAMORPHISM

Criteria used in determining metamorphic facies and grade in this study are principally those outlined by Winkler (1967, 1976). Additional criteria for mafic rocks were taken from Froese (1973) and for iron formations from James (1955). Particular attention was paid to amphibole colour and type, textural relationships of coexisting pyroxenes and amphiboles, and to varieties of feldspar present.

The region was affected by at least two Archean metamorphic episodes, but from the present study their mutual relationship is uncertain. Most of the Archean rocks are in the amphibolite facies; because of insufficient data from both field and petrographic studies, subdivision of the facies into upper and lower amphibolite categories was not possible. A large area in the southeast part of the region is underlain by granulite facies rocks, whereas large areas in the north and west are underlain by both amphibolite and granulite facies rocks. In the latter case, there is evidence that amphibolite metamorphism post dated granulite metamorphism. The widespread development of chlorite, epidote-group minerals, albite, and myrmekite, suggests a greenschist to subgreenschist facies retrogression; however, large areas underlain only by greenschist rocks are rare.

Representative mineral associations are listed in Table 1. Pressure-temperature indicator assemblages are scarce. Central Ungava (northern portion of Leaf River area of Stevenson (1968), south of Cape Smith Belt) contains a relatively high proportion of pelitic rocks; supracrustal rocks east of James Bay (south of 54°N) also contain useful metasedimentary associations. The granulite facies terrane in the southwest -- Ashuanipi complex (Baragar, 1967) -- contains pelitic associations in paragneiss.

Amphibolite facies associations in quartz-bearing pelitic rocks include cordierite, garnet, staurolite, muscovite, biotite, and the aluminum silicates sillimanite and andalusite, but not kyanite; K feldspar is generally absent. Metamorphism at relatively low pressures and temperatures within the amphibolite facies is indicated. The Ashuanipi complex comprises masses of garnet-cordierite-orthopyroxene-biotite-perthite/antiperthite gneiss, perhaps indicative of relatively high temperatures in granulite facies. Occurrences of garnet with hornblende in mafic rocks along the Archean craton boundaries and also near the Ashuanipi complex suggest relatively high pressure at amphibolite grade. Elsewhere, however, the association cummingtonite-hornblende-plagioclase indicates relatively low pressure during amphibolite facies (Shidô, 1958; Froese, 1973). Granulite facies rocks which have been overprinted by a lower grade of metamorphism commonly contain prehnite-biotite intergrowths (Phillips and Rickwood, 1975; Field, 1976; Rickwood and Phillips, 1976); these intergrowths are also common in amphibolite facies rocks overprinted by greenschist facies.

Effects of Proterozoic metamorphism may be recognized to varying degrees in all rocks surrounding the Archean craton. Prehnite-pumpellyite assemblages occur in metabasalts near Cape Smith, indicating at least local subgreenschist facies conditions. Effects and extent of metamorphism in the Otish Mountains, Nastapoka Arc, and in the Sakami Formation outliers are difficult to assess because of the lack of diagnostic mineral associations. Quartz and sericite/muscovite are predominant in psammitic-pelitic rocks of the Proterozoic formations. One occurrence of andalusite-muscovite-quartz occurs apparently in a Sakami Formation outlier, indicating upper greenschist to lower amphibolite facies conditions. A well-defined greenschist facies overprint on amphibolite facies along the Grenville-Superior boundary may be Grenvillian in age.

Precambrian diabase/gabbro dykes throughout New Quebec, variably metamorphosed probably at subgreenschist to greenschist grade, contain zeolites, prehnite, fibrous amphiboles, and pumpellyite.

Evidence of shock metamorphism has been found or postulated at four small areas in New Quebec (Fryer and Titulaer, 1972). These are New Quebec crater, Lac Couture, and Clearwater Lake (east and west); the Manicouagan ring structure is just outside the area studied. Sanidinite facies contact metamorphic conditions are indicated by the occurrence of spurrite in limestone digested by lava at west Clearwater Lake. There is local extensive zeolitic alteration in the lava, and the shock metamorphism is clearly indicated by maskelynite in granulite facies gneiss.

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ARCHEAN REGIONAL METAMORPHISM IN THE SLAVE STRUCTURAL PROVINCE – A NEW PERSPECTIVE ON SOME OLD ROCKS

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Thompson, P.H., Archean regional metamorphism in the Slave Structural Province — A new perspective on some old rocks; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 85-102, 1978.

Abstract

The distribution of Archean regional metamorphism is a record of a critical time in the history of the Slave Structural Province. A patchy pattern of greenschist, lower amphibolite, and upper amphibolite facies metamorphism has been superimposed on the rocks of the Yellowknife Supergroup and underlying granitic basement subsequent to the main phases of folding during a tectonic event about 2600 Ma ago. Throughout the area (190 000 km^2) the cordierite-biotite-andalusite-sillimanite (staurolite) assemblages are typical of those produced by low pressure regional metamorphism of the Pyreneean type. At four localities in the northeastern part of the province, kyanite occurs with cordierite, staurolite, and/or sillimanite, indicating that these rocks were in the stability field of kyanite for part of their metamorphic history.

The patchy variation of metamorphic grade is the product of differential erosion of an irregular thermal topography (thermal ridges or domes and depressions) preserved in the rocks as the isograd pattern. Plotted on a P-T diagram, P-T curves derived from traces across the erosion surface imply postmetamorphism uplift ranging from less than 5 km in low grade areas to 12-15 km in high grade areas. Seismic data from the southern part of the Slave Province provide an estimate of 34 km for the present crustal thickness. As it is unlikely that tens of kilometres of sialic rocks have been added since the 2600 Ma tectonic event, the sialic crust may have been as much as 40-50 km thick when metamorphism attained a maximum. Crustal thickening is attributed to horizontal compression rather than addition of material from the mantle.

Résumé

La distribution du métamorphisme régional archéen est le vestige d'une époque critique de l'histoire de la province structurale des Esclaves. Un métamorphisme sporadique, caractérisé par des faciès des schistes verts, roches du supergroupe de Yellowknife et au soubassement granitique sous-jacent, à la suite des principales phases de plissements, pendant un épisode tectonique qui remonte à environ 2 600 Ma. Dans toute la région (190 000 km^2) les assemblages à cordiérite-biotite-andalousite-sillimanite (staurolite) sont typiques du métamorphisme régional de type pyrénéen produit à des pressions faibles. Dans quatre localités de la partie nord-est de la province, la cyanite est associée à la cordiérite, à la staurolite, et parfois à la sillimanite, ce qui indique que ces roches se sont trouvées dans le champ de stabilité de la cyanite pendant une partie de leur évolution métamorphique.

Les variations sporadiques de l'intensité du métamorphisme résultent de l'érosion différentielle d'une topographie thermique irrégulière (crêtes ou dômes et dépressions thermiques), préservée dans les roches par les surfaces d'isogrades. Représentées sur un diagramme P-T, les courbes P-T obtenues à partir des traces qui parcourent la surface d'érosion indiquent un soulèvement post-métamorphique qui varie entre moins de 5 km dans les zones de degré métamorphique faible et 12 à 15 km dans les zones de degré métamorphique élevé. Les données sismiques obtenues dans la partie sud de la province des Esclaves nous permettent d'estimer à 34 km l'épaisseur actuelle de la croûte. Comme il est peu probable que des dizaines de kilomètres de roches sialiques aient été ajoutées depuis l'événement tectonique survenu il y a 2 600 Ma, il est probable que la croûte sialique a atteint 40 à 50 km d'épaisseur pendant la culmination du métamorphisme. L'épaississement de la croûte est attribué à une compression horizontale, plutôt qu'au rajout de matériaux provenant du manteau.

INTRODUCTION

The Slave Structural Province (~190 000 km^2) in the northwestern part of the Canadian Shield is underlain by rocks that were deformed, metamorphosed, and intruded by plutonic rocks during a major tectonic event approximately 2600 Ma ago. The compilation of a metamorphic map in conjunction with the Geological Survey of Canada project Metamorphism in the Canadian Shield provided an excellent opportunity to consider, for the first time on a regional scale, Archean metamorphism in the Slave Province. Knowledge of the extent and intensity of metamorphism and its relation to deformation and plutonic activity is fundamental to an understanding of the geology of this province. In this paper, the variation of metamorphic grade as outlined by the isograd pattern is used as a basis for estimating the range of geothermal gradients and, thereby, the distribution of pressure and temperature in the crust when the isograds formed. This information in the context of other aspects of the geology provides a new perspective on the 2600 Ma tectonic event in the Slave Province.

GENERAL GEOLOGY

McGlynn and Henderson (1970, 1972) and McGlynn (1977) provide an excellent summary of the general geology of the Slave Province. Present knowledge is based on regional mapping (Barnes and Lord, 1954; Bostock, 1976; Folinsbee, 1949, 1952; Fortier, 1949; Fraser et al., 1960; Fraser, 1964, 1969; J.B. Henderson, 1976; J.F. Henderson, 1939, 1941a,b,c, 1944; Jollife, 1942, 1946; Lord, 1942; McGlynn and Ross, 1962; Moore et al., 1951; Padgham, 1974; Padgham et al., 1974; Ross, 1966; Tremblay et al., 1947, 1954; Tremblay, 1971, 1976; Wilson and Lord, 1942; Wright, 1954, 1967; Yardley, 1949) and detailed studies (see below) covering small areas at a number of localities. To the north

[1] Geological Survey of Canada

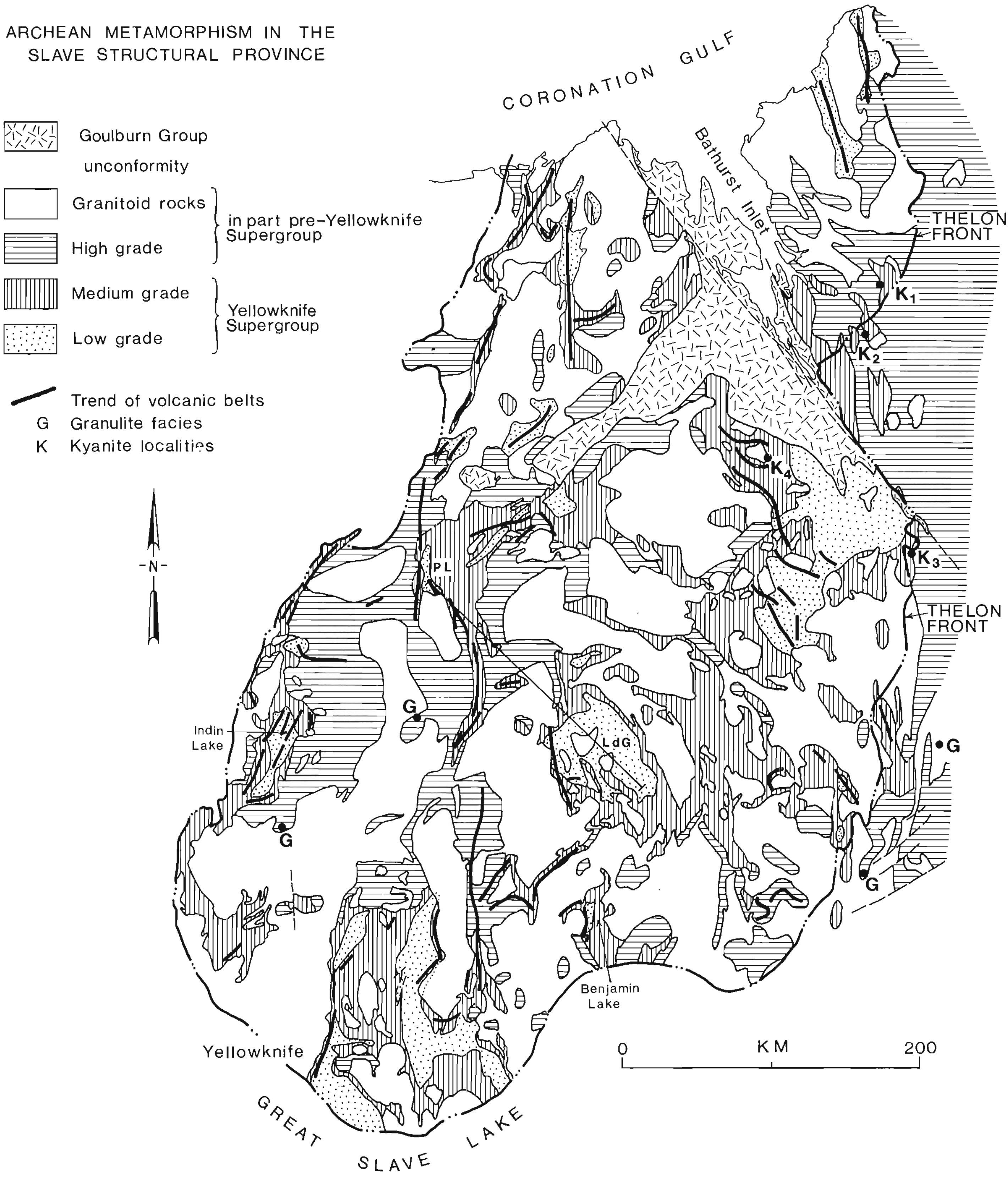


Figure 1. Metamorphic map of the Slave Province. The geology of the Churchill Province east of the Thelon Front was compiled from Fraser (1964, 1968, 1972), Wright (1967), and Tremblay (1971). The location of a cross-section between Point Lake (PL) and Lac de Gras (LdG) is indicated.

and south (Fig. 1) the Slave Structural Province (Stockwell, 1964) is bounded by unmetamorphosed rocks of Aphebian to Paleozoic age and by Coronation Gulf and Great Slave Lake. The Goulburn Group is a remnant of the Aphebian cover preserved inside the province. To the west, the boundary with the Bear Province is marked by an unconformity at the base of the Aphebian Snare and Epworth groups and a tectonic-metamorphic discontinuity related to a younger tectonic event. The Thelon Front has been taken as the boundary with the Churchill Structural Province to the east (Wright, 1967; Fraser, 1964, 1968).

The Slave Province (Fig. 1) is composed of Archean supracrustal rocks (∽35%) designated the Yellowknife Supergroup by Henderson (1970b) and gneiss-migmatite-granitoid complexes (∽65%). At different localities the latter are the metamorphosed equivalents of the supracrustal rocks, basement to these rocks, and intrusive into them. The lithologies, structural style, and type of metamorphism are quite uniform. The fact that 85 per cent of the Yellowknife Supergroup is metasedimentary rock (McGlynn and Henderson, 1972) distinguishes the Slave Province from other Archean terranes such as the Superior Structural Province where volcanic rocks predominate. Yellowknife volcanic rocks range from basalt to rhyolite with dacite-rhyolite predominant in some volcanic belts (Baragar, 1966; Padgham, 1974; Lambert, 1976; McGlynn, 1977). In areas of low grade metamorphism the most common metasedimentary rocks, metagreywacke and slate, exhibit primary sedimentary structures characteristic of turbidites (Henderson, 1970a; Henderson, 1975a,b). Locally, metamorphosed conglomerate, carbonates, and tuffaceous rocks occur at the contact between metavolcanic rocks and overlying metasediments. Based on measurements made in low grade, relatively undeformed rocks, the thickness of the supracrustal sequence is variable but less than 5-10 km (McGlynn and Henderson, 1972; Henderson, 1975a,b; Tremblay, 1976). This may be a minimum estimate; the original thickness has not been measured in medium to high grade metamorphic rocks where deformation was more intense.

Present knowledge of the gneiss-migmatite-granitoid complexes suggests a complicated history. In the past gneisses and migmatites have been mapped as derivatives of the Yellowknife Supergroup. Recent work suggests that gneissic rocks west of the supracrustals at Point Lake (Bostock, 1976; Henderson and Easton, 1977) and near Ross Lake (70 km east of Yellowknife) (Baragar, 1966; Davidson, 1972) are basement to the supracrustal rocks. Relatively homogeneous granitoid rocks, mainly quartz diorite, granodiorite, and quartz monzonite, occupy about 30 per cent of the Slave Province. Generally, these rocks were considered to be the youngest rocks exposed. Detailed investigations have shown that in many areas the plutonic rocks postdate the regional metamorphism of the sedimentary and volcanic rocks they intrude (Henderson, 1943; Moore, 1956; Ross, 1966; Davidson, 1967; Kretz, 1968; Tremblay, 1976). At Point Lake (Stockwell, 1933; Henderson and Easton, 1977), Benjamin Lake (Heywood and Davidson, 1969), Indin Lake (Frith et al., 1977), and in the Cameron River area (Baragar, 1966; Davidson, 1972), however, some of the granitoid rocks are basement to the Yellowknife Supergroup.

STRUCTURAL GEOLOGY AND GEOCHRONOLOGY

To integrate regional metamorphism into the geological history of the Slave Province the structural setting and timing of deformation with respect to metamorphism is considered. Only a few detailed studies of structural geology have been done in the Slave Structural Province (Henderson, 1943; Tremblay, 1952, 1976; Ross, 1966; Ross and McGlynn, 1965; Davidson, 1967; Smith, 1966; Ramsay, 1973a; Fyson, 1975). The summary presented here is based on these works and on maps and reports referred to previously.

Structural geometry can be attributed to deformation associated with the 2600 Ma tectonic event. However, structures in gneissic or granitoid rocks interpreted to be basement to the Yellowknife Supergroup may be much older. Post-Archean folding of Archean rocks occurred along the boundary with the Bear Province after deposition of the unconformably overlying Snare and Epworth groups (Ross, 1966; Ross and McGlynn, 1965; Hoffman, 1971). Faulting and mylonitization along the eastern boundary and near Bathurst Inlet postdate the Aphebian Goulburn Group (Wright, 1967; Tremblay, 1971) but a long history of movement in this area is suggested by petrographic data, indicating that some rocks were sheared prior to the main Archean metamorphism (this study). Widespread post-Archean fracturing accompanied by intrusion of diabase sills and dyke swarms occurred several times after the tectonic activity in the Slave Province had ended.

The extent to which deformation related to the 2600 Ma event affected volcanic and plutonic rocks has not been studied in detail. In most cases, metavolcanic rocks adjacent to complexly folded metasediments form a homoclinal sequence; but at some localities they are isoclinally folded (Lambert, 1977) and contain flattened or elongated pillows (Heywood and Davidson, 1969). The structural relations of plutonic rocks in the Slave Province are complex. At Benjamin Lake (Heywood and Davidson, 1969) plutonic basement rocks were deformed internally and, at the same time, acted as buttresses around which supracrustal rocks were deformed. On the other hand, the youngest plutons related to the 2600 Ma event are massive and clearly intrude and deform the previously metamorphosed and deformed country rock. Aside from measurements of foliation and compositional layering in gneiss-migmatite complexes little is known about the deformation of these rocks. The metasedimentary rocks provide the most complete record of deformation. Although structural style is similar throughout the province, the complexity of the fold pattern and the degree of deformation, e.g. tightness of folds, vary considerably. Data are limited but, in most areas, the structural geometry can be represented by a three-phase sequence, two major periods of folding (D_1 and D_2) followed by deformation (D_3) related to the intrusion of postmetamorphism plutons.

Reversals in facing direction obtained from well-preserved sedimentary structures are the main evidence for D_1 folding throughout the province (e.g., Henderson, 1943; Folinsbee, 1949; Tremblay, 1952; Davidson, 1967; Fraser, 1969; Bostock, 1976). Folds, where preserved, are isoclinal with steeply dipping axial surfaces and variably plunging hinge lines. Fold hinges and entire limbs are commonly sheared out (Henderson, 1943; Davidson, 1967; Tremblay, 1952, 1976; Bostock, 1976). Cleavage associated with these folds is sparsely developed and difficult to distinguish from a preferred orientation developed during sedimentation and compaction. The presence of basement blocks during D_1 probably produced a variety of orientations for D_1 structures (e.g. Heywood and Davidson, 1969). Greenschist facies metamorphism prevailed during D_1 (Davidson, 1967; Fyson, 1975).

During the early part of the second phase of deformation D_1 isoclines were tightened or refolded. Second folds have steeply dipping axial surfaces with the plunge of fold hinge lines dependent to some extent on the original orientation of the folded surface. A prominent, steeply dipping, north-trending schistosity (S_2) is axial planar to these folds and cuts across both limbs of D_1 isoclines without changing orientation (Henderson, 1943; Tremblay, 1952; Davidson, 1967). At several localities interference patterns formed by superposition of D_1 and D_2 structures have been observed (Henderson, 1943; Fortier, 1946, 1947; Fyson, 1975); however, wherever the original orientation of D_1 axial surfaces was parallel to the D_2 axial surfaces it is difficult to distinguish two phases of folding. Porphyroblasts of

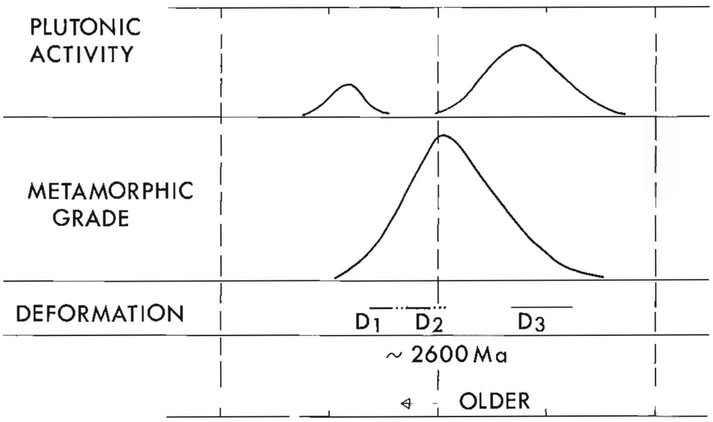


Figure 2. Time sequence of the main stages of the 2600 Ma tectonic event in the Slave Province. The early peak of plutonic activity is related to formation of volcanic rocks of the Yellowknife Supergroup.

cordierite and andalusite overgrow the mica schistosity (S_2) both randomly and with a preferred orientation that may or may not be parallel to the schistosity (e.g., Bostock, 1976; Tremblay, 1952, 1976).

Evidence for a third phase of deformation in the form of the crenulation of S_2 tends to be spatially related to plutonic bodies that postdate the peak of metamorphism (Davidson, 1967; Frith et al., 1977). An accessible example of the crenulation of S_2 is located 40 km east of Yellowknife on the road to Tibbett Lake (Cameron River Bridge) where the growth of new chlorite and retrogression of cordierite at the time of crenulation supports the conclusions of previous workers that metamorphic grade was decreasing during D_3.

The limited amount of structural data from the Slave Province permits the following conclusions. There is widespread evidence, mainly in the supracrustal rocks but also in the basement, of a moderate to high degree of deformation that can be attributed to the 2600 Ma tectonic event. Textural relations and the fact that isograds transect S_2 and the axial surfaces of major folds indicate that the peak of metamorphism occurred after the main phases of folding in the Slave Province (Henderson, 1943; Tremblay, 1952; Davidson, 1967; Bostock, 1976). There is a predominantly northerly trend (northwest to northeast) to volcanic belts in the province (Fig. 1) and to the main schistosity recorded by previous workers (see references above) which may reflect an east-west to northeast-southwest orientation for axes of compression during D_1 and D_2. It is possible that large scale horizontal crustal shortening (variable but as much as or more than 50 per cent) produced the observed deformation.

Detailed geochronological studies are limited to the Yellowknife (Green and Baadsgaard, 1971) and Indin Lake (Frith et al., 1977) areas. Green and Baadsgaard concluded that the history of the Slave tectonic event began with volcanism at 2700-2600 Ma and ended with the intrusion of potassic granitoid rocks at 2575 Ma. The discovery of 3000 Ma ages for granitic basement at Yellowknife (Nikic et al., 1975), near Indin Lake (Frith et al., 1977), and a preliminary zircon age from granitic basement at Point Lake (J.B. Henderson, pers. comm., 1977) indicates that the rocks of the Slave Province represent a longer period than was deduced from earlier age determinations. The exact timing of deformation, metamorphism, and plutonic activity is not known but results of detailed mapping by Henderson (1939, 1941a,b,c, 1943), Tremblay (1952, 1976), and Davidson (1967) support the sequence of events in Figure 2.

It is not known whether each stage of the geological development, i.e., volcanism, subsidence, sedimentation, deformation, metamorphism, and plutonic activity occurred simultaneously throughout the province or in a time sequence migrating across the province. It may be assumed for the present that each stage occurred at approximately the same time throughout the province.

METAMORPHISM

The metamorphic map (Fig. 1) is a record of regional metamorphism which reached a maximum during the 2600 Ma tectonic event. The variations in metamorphic grade reflect the distribution of pressure and temperature when the pattern formed as well as the amount of subsequent uplift and erosion. To obtain metamorphic P-T distributions from a metamorphic map requires a number of steps:

1) from isograds to mineral equilibria or reactions expressed in terms of pressure, temperature, and the activities of components such as water. Isograds may be represented by discontinuous mineral reactions (a line on a P-T diagram) or continuous mineral reactions (a P-T band). The significance of these types of reactions and the problem of the activity of water are discussed in a subsequent section of the paper.
2) from mineral reactions on a P-T diagram to erosion surface P-T curves and limiting metamorphic geothermal gradients.
3) from erosion surface P-T curves and geothermal gradients to a P-T profile consistent with these observations.

The application of this approach to the Slave Province involves a certain amount of conjecture and interpretation because much of the mapping is at a scale of 1:250 000 and most of the northern region is known only from helicopter reconnaissance.

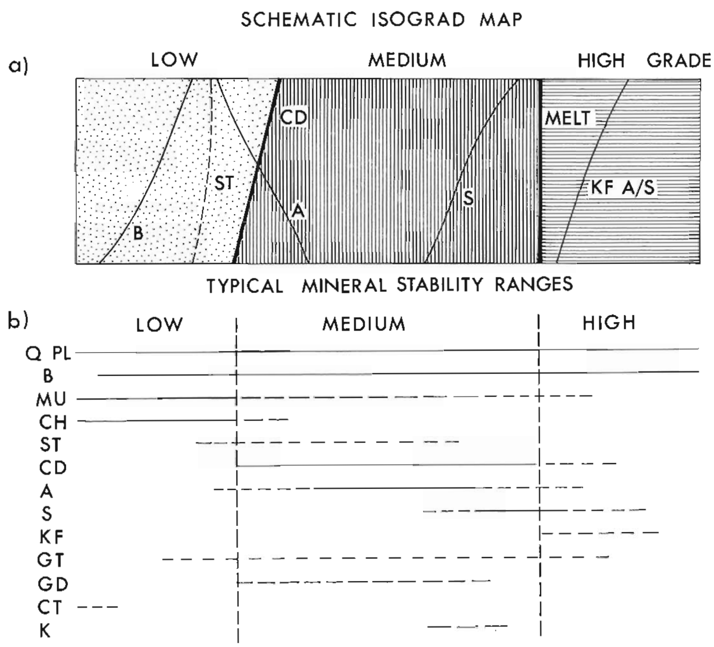


Figure 3. a) Schematic composite map of isograds in the Slave Province.

b) Typical stability ranges for metamorphic minerals. Mineral abbreviations:

Q – quartz	A – andalusite
PL – plagioclase	S – sillimanite
B – biotite	K – kyanite
Mu – muscovite	KF – k-feldspar
CH – chlorite	GT – garnet
ST – Staurolite	GD – gedrite
CD – cordierite	CT – chloritoid

Metamorphic Map

On the basis of McGlynn's (1977) geological compilation map, reports and maps previously referred to, detailed studies of small areas (Denton, 1940; Wright, 1950; Smith, 1966; Davidson, 1967; Kretz, 1968, 1973; Ramsay, 1973a; Kamineni, 1973, 1975), and examination of hundreds of thin sections from the Geological Survey collection, three metamorphic zones are distinguished on Figure 1 with low, medium, and high grade corresponding to greenschist, lower and upper amphibolite facies, respectively. The low/medium grade boundary marks the transition from slate-metagreywacke to knotted schist in metasedimentary rocks of the Yellowknife Supergroup. In most cases the medium/high grade boundary separates knotted schist from gneiss and migmatites (rocks with 25-75 per cent magmatic aspect). Some small patches of migmatite may be the product of the intrusion of postmetamorphism plutons. McGlynn and Henderson (Fig. 3, 1970) imply that some of the high grade rocks may be basement to the Yellowknife supracrustal rocks; this is also suggested by field relations west of the low grade rocks at Point Lake (Bostock, 1976; Henderson and Easton, 1977). Elsewhere in the Slave Province, previous workers (e.g., Tremblay et al., 1947, 1954; Moore et al., 1951; Folinsbee, 1952; Fraser, 1969; Heywood and Davidson, 1969; Bostock, 1976) mapped transitions from knotted schist to migmatite and considered the migmatites to have been derived from supracrustal rocks. The percentage of basement rocks in gneissic and migmatitic terranes is one of the major unsolved problems of the Slave Province but it is reasonable to suggest that most of the high grade rocks are derived from supracrustal rocks. The origin of the gneisses and migmatites is irrelevant in establishing the metamorphic pattern; low and medium grade rocks are sufficient. East of the Thelon Front (Fig. 1), however, large tracts of gneissic rocks may record a pre-2600 Ma metamorphic event.

Low grade zones occur randomly across the province within areas of medium grade that in turn are surrounded by high grade rocks (Fig. 1). Granulite facies rocks are known from two localities in the province (Folinsbee, 1940; Fraser, 1969). The "patchy" metamorphic pattern appears to be superimposed on the main structural trends as outlined by belts of metavolcanic and metasedimentary rocks, an observation in agreement with the conclusion based on textural data and detailed mapping that the metamorphism outlasted the main phases of deformation.

The predominance of granitoid rocks in medium and high grade zones (Fig. 1) has been interpreted as an indication that the plutonic rocks were the source of heat for the metamorphism of the supracrustals (e.g., Henderson, 1943; Folinsbee, 1949; Lambert, 1976). It can be seen, however, that plutons truncate metamorphic zone boundaries and intrude low grade rocks. In the northern part of the province large masses of predominantly low grade rocks are surrounded by granitoid rocks which may represent either an earlier basement or later, relatively cold intrusions. In the Benjamin Lake area Davidson (1967) suggested that the granitoid rocks were produced at depth by the same thermal conditions that metamorphosed the supracrustal rocks and that the magma subsequently intruded the metamorphic rocks; this origin is consistent with the pattern of metamorphic zones throughout the province.

Mineral Assemblages

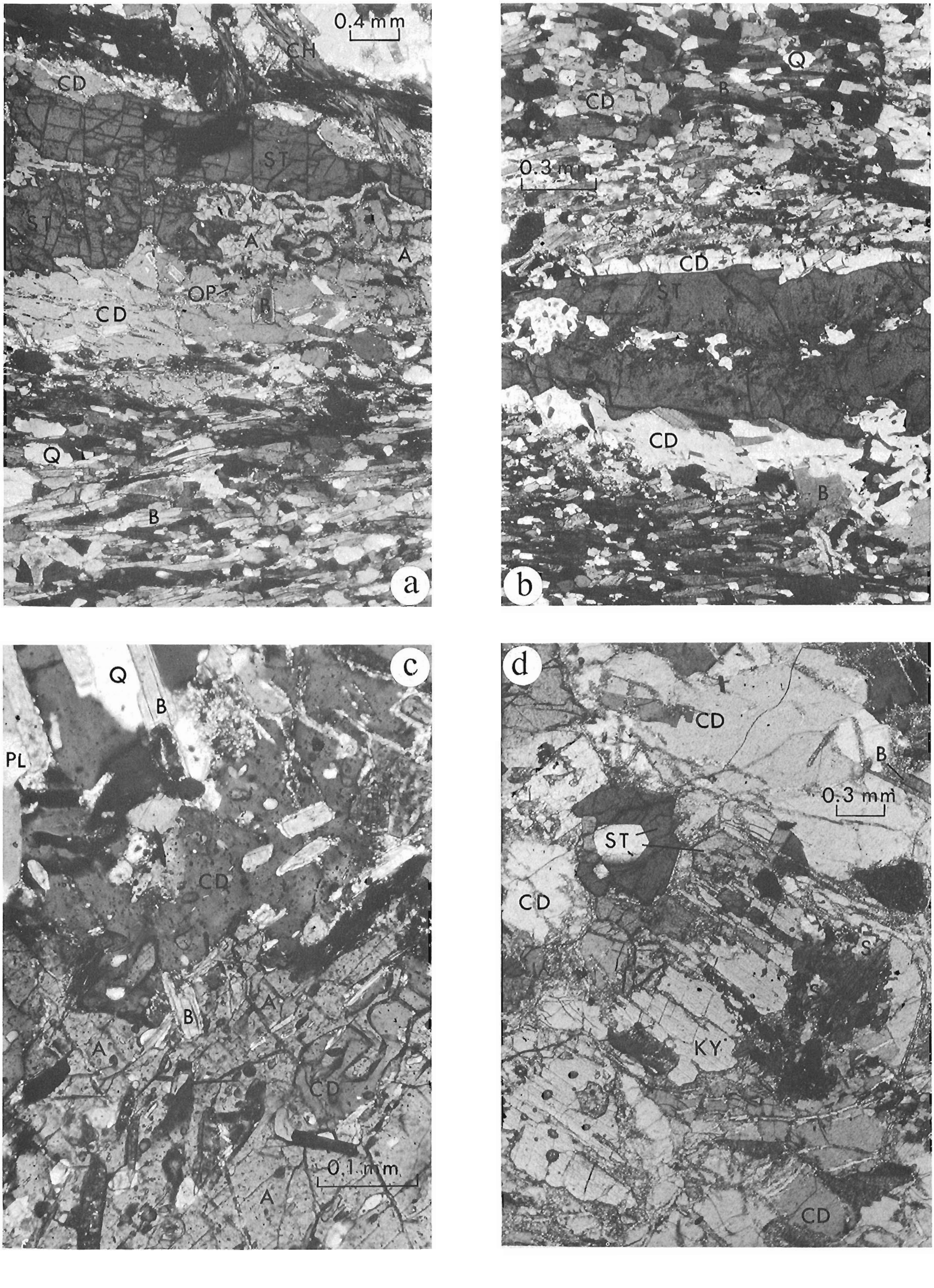
The uniform composition of Yellowknife metasediments in the Slave Province (McGlynn and Henderson, 1972; Henderson, 1975b) is evident from the overall similarity of mineral assemblages in each of the metamorphic zones (Fig. 3). The assemblages indicate that low pressure metamorphism of the Pyreneean type (Hietanen, 1967) prevailed when the metamorphic maximum was attained. The relative positions of the first appearance of a mineral and the textural relations vary somewhat across the province but, in spite of the variable density of observations, some generalizations can be made. Chloritoid is present only in low grade rocks near Yellowknife (Folinsbee, 1942; Henderson, 1975a). Garnet occurs sporadically in rocks of different metamorphic grade and probably reflects a locally high concentration of Fe, Mn, and/or Ca. Cordierite is the most common constituent of the knots in the knotted schist; it may be accompanied by andalusite. The relative rarity of staurolite can be related to the typically low Fe content of Slave metagreywackes (see below). It commonly occurs in the cores of cordierite or andalusite grains (Fig. 4). In a small area east of Yellowknife, Kamineni (1975) outlined a gedrite zone in medium grade rocks; elsewhere data are insufficient to define the distribution of this mineral. Andalusite, as porphyroblasts or in quartz segregations, is the most common polymorph of Al_2SiO_5. At some localities it occurs inside cordierite as well (Fig. 4). Fibrolitic sillimanite is present with cordierite and andalusite, in the upper part of the medium grade zone. K-feldspar is relatively rare in nonmigmatized rocks. In part, this may be due to the physical conditions of metamorphism but the disappearance of muscovite from knotted schists at medium grade, leaving the common assemblage quartz-plagioclase-biotite-cordierite-andalusite-sillimanite, precludes the formation of K-feldspar by the reaction of muscovite with quartz. K-feldspar coexists with cordierite and sillimanite south of Indin Lake (Folinsbee, 1942; Lord, 1942), andalusite and sillimanite ± cordierite near Point Lake (Bostock, 1976), and with cordierite (sillimanite?) in the migmatites 120 km north of Yellowknife (Moore et al., 1951).

Kyanite is limited to sillimanite-bearing rocks in the northeastern part of the province (Fig. 1). The localities east (K_1) and southeast (K_3) of the Bathurst Inlet are on the border between an area characterized by the assemblage cordierite-sillimanite-biotite ± staurolite, ± andalusite and a large area to the east where few data are available but in which granulites were recognized. The kyanite and sillimanite in sample K_3 have overgrown a micaceous matrix that appears to be in a recrystallized shear zone in a rock that was originally gneissic or granitoid. A thin section made from a sample supplied by R.A. Frith (K_4) revealed kyanite coexisting with staurolite and sillimanite inside a large grain of cordierite (Fig. 4). This texture does not necessarily imply a decrease in pressure after the formation of kyanite. Although relatively rare, the presence of this high pressure polymorph of Al_2SiO_5 has been reported from predominantly low pressure terranes in the Archean of southern Africa (Saggerson and Turner, 1972, 1976), Australia (Binns et al., 1976), Canada (Schau, 1978; Jolly, 1978) and Greenland (Bridgwater, 1974) and in the low pressure Hercynian rocks of the eastern Pyrenees (Fonteilles et al., 1964).

Isograds

Isograd is used in this paper as a field term for a line marking the appearance or disappearance of a mineral or mineral assemblage in rocks of similar composition. Where detailed information is available, the relationships between isograds and isotherms, isobars, or specific mineral reactions are considered. Even without petrographic data, however, the isograd pattern as mapped in the field significantly limits the range of possible metamorphic P-T conditions.

The knotted schist (cordierite, cordierite-andalusite) isograd and the knotted schist/migmatite transition or melting (25-75 per cent) isograd have been included on geological maps in the Slave Province since the late 1930's (Henderson, 1939; Jollife, 1942). The biotite, andalusite, and sillimanite isograds were located on maps by Tremblay et al. (1947, 1954), Wright (1954), Heywood and Davidson (1969),


0.4 mm
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Henderson and Easton (1977), and Bostock (1976). Bostock (1976) also describes a K-feldspar-aluminosilicate isograd on the high grade side of the melting isograd east of Point Lake (Fig. 1).

The variable spatial relations of the isograds in the Slave Province are illustrated schematically in Figure 3. It is necessary to interpret the isograds in terms of mineral reactions to determine the extent to which the variation is the result of differences in rock composition, pressure, temperature gradients, or some combination of these parameters.

Mineral Reactions and Rock Composition

The change in mineral assemblage marking an isograd may be represented as a specific reaction among the minerals (Chinner, 1966; Carmichael, 1970). On a triangular phase diagram mineral reactions correspond either to a discontinuous change requiring an exchange of tie-lines or to a continuous change involving the movement of a subtriangle across the face of the main triangle (Fig. 5). Isograds that can be represented as discontinuous reactions are preferable because on a P-T diagram such reactions plot as lines that apply to a wide range of composition. A P-T curve representing a continuous reaction applies only to a specific composition; a different composition requires a different P-T curve. If the range of rock composition is restricted, however, the family of P-T curves will plot within a narrow band. In any case, the mineralogical changes in the rocks determine which reaction is most applicable. In the Slave Province, the composition of the metagreywackes is such that, in many cases, continuous mineral reactions provide the most accurate representation of the cordierite, andalusite, and sillimanite isograds.

Bulk compositions of the Yellowknife Supergroup metasediments can be plotted on an AFM projection (Fig. 6) that is valid in the presence of quartz, oligoclase, muscovite, magnetite, and ilmenite. The diagram is similar to the one used by Reinhardt (1968), except that compositions have been projected through muscovite rather than K-feldspar. The diagram emphasizes Henderson's (1975a,b) conclusion that the metagreywackes from the Yellowknife area are somewhat

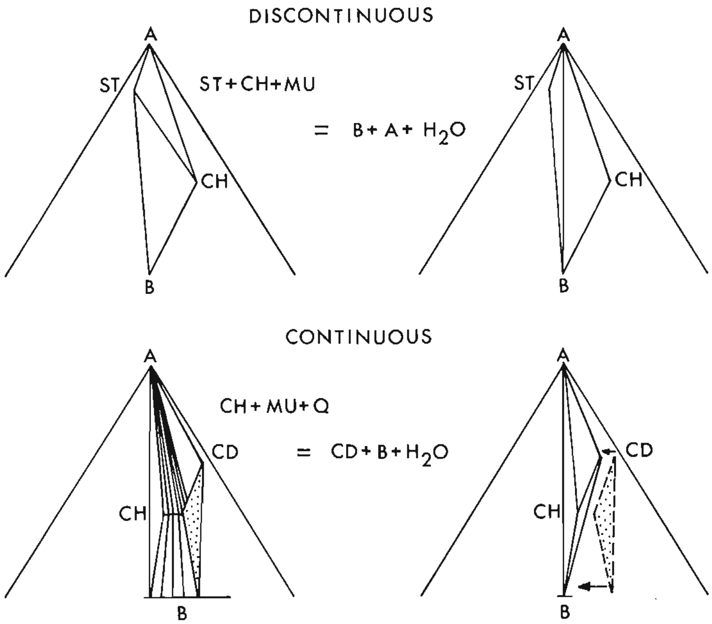


Figure 5. Discontinuous and continuous reactions on an AFM projection. Quartz and muscovite present. The activity of water is constant. Mineral abbreviations as in Figure 3.

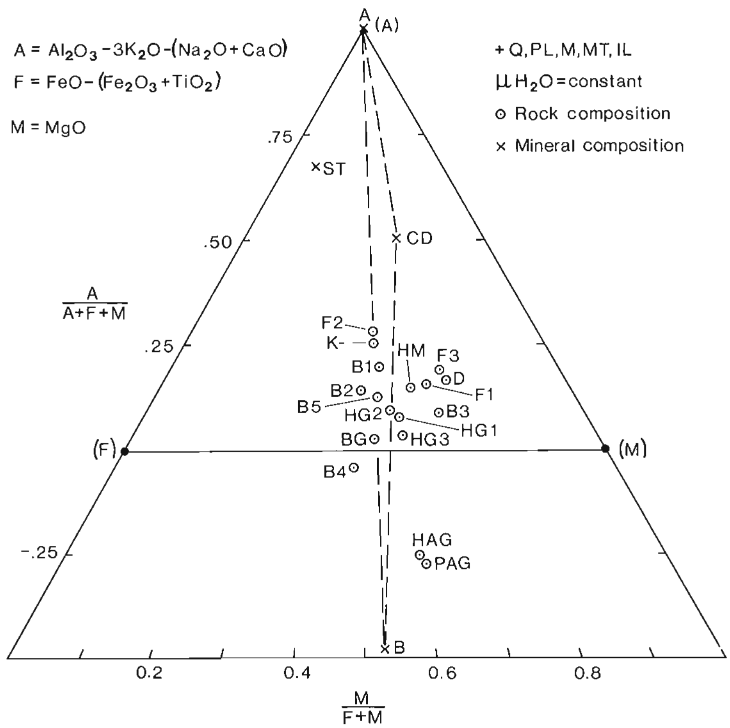


Figure 6. AFM diagram obtained by projection through quartz, muscovite, oligoclase, magnetite, and ilmenite.

Rock compositions (Fe_2O_3 analyzed) of metasedimentary rocks from the Slave Province:

Burwash Formation metagreywackes (Yellowknife area), HG (Henderson, 1975a,b), BG (Boyle, 1961);

average greywacke, PAG (Pettijohn, 1957), HAG (Archean, Henderson, 1975a,b);

metamudstone, HM (Henderson, 1975a);

knotted schists, D (Davidson, 1967), F (Folinsbee, 1942), K (Kamineni, 1973). Coexisting biotite-cordierite from Kamineni (1973).

The compositions are plotted on a Thompson (1957) AFM projection following the method of Reinhardt (1968). Mineral abbreviations as in Figure 3 with:
Mt – magnetite, (A) – andalusite,
IL – ilmenite, M – muscovite.

Figure 4 (opposite)

Overgrowth textures in knotted schists:

a) staurolite (ST) – andalusite (A) overgrown by cordierite (CD) from the Indin Lake area (W-165-47; Wright, 1950) (GSC 203326);

b) staurolite (ST) rimmed by cordierite (CD) from Yellowknife area (HBA-k.496-2-71; J.B. Henderson, pers. comm., 1976) (GSC 203326-A);

c) andalusite (A) rimmed by cordierite (CD), same sample as b (GSC 203326-B);

d) kyanite (K), staurolite (ST), and sillimanite (S) included in cordierite (CD) (k4 Fig. 1, collected by R.A. Frith) (GSC 203326-C).

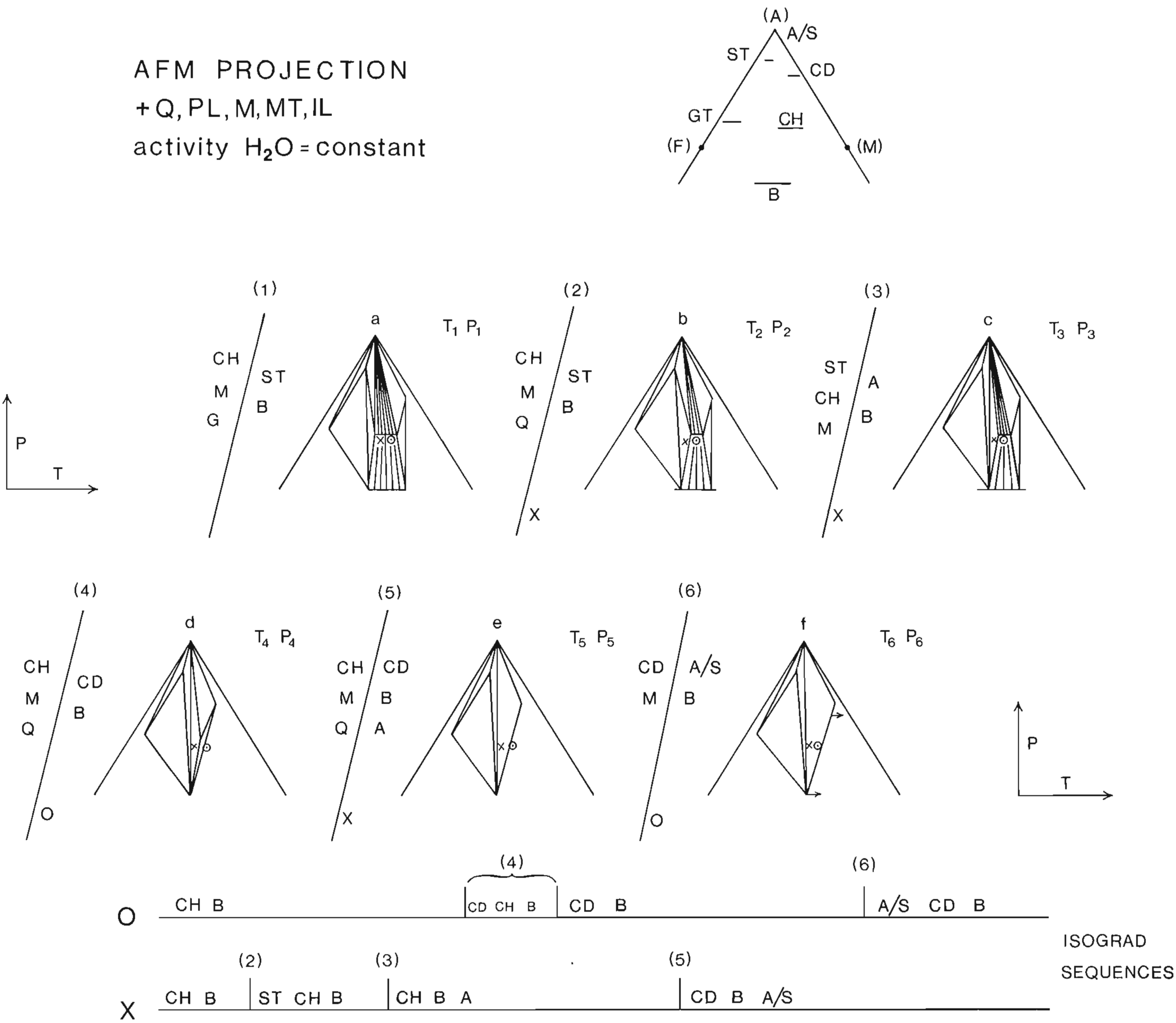


Figure 7. A series of AFM projections (a-f) and mineral reactions (1-6) are plotted with temperature and pressure increasing. The continuous reactions 4 and 6 apply only to the rock composition O and reaction 2 only to composition X. Reactions 1, 3, and 5 are discontinuous reactions. Mineral zonation and isograds associated with each composition are included. Mineral abbreviations are the same as on Figure 3 with MT-magnetite, IL-ilmenite.

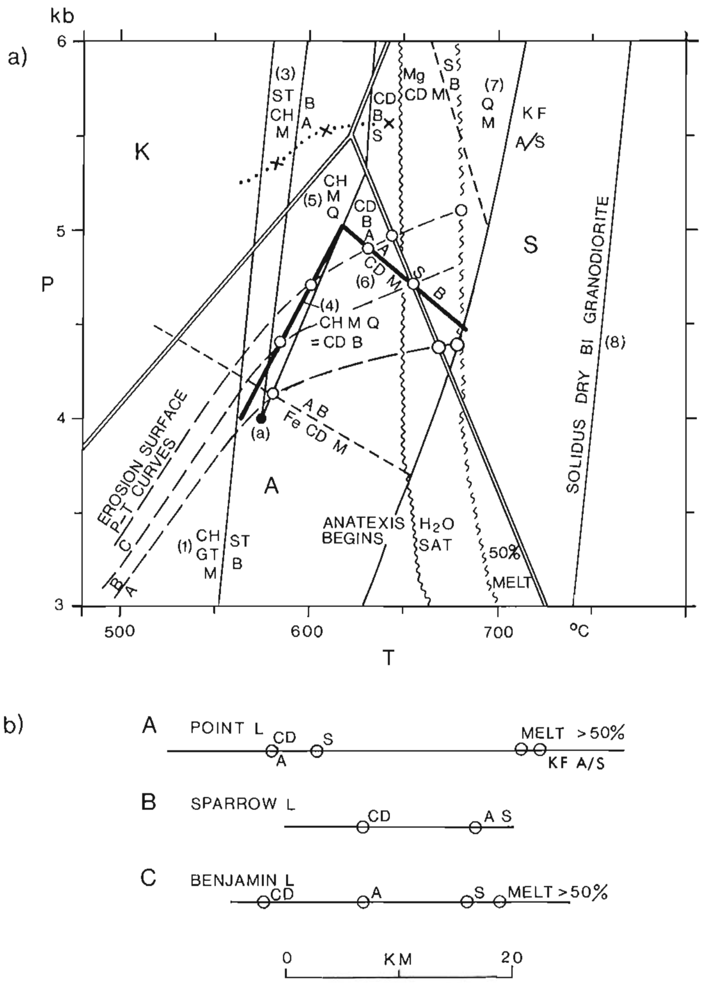
more Fe-rich and more aluminous than Pettijohn's (1957) average Archean greywacke. That many of the rocks plot to greywacke. That many of the rocks plot to the right of a biotite-cordierite tie-line from a typical knotted schist helps to explain the predominance of cordierite over staurolite in medium grade rocks of the Slave Province. A series of AFM projections (Fig. 7) illustrates the significance of these compositions with respect to mineral reactions. With increasing metamorphic grade, rock O, a typical Slave greywacke (quartz-K mica-plagioclase-chlorite-biotite-Fe-Ti oxides), is not affected by discontinuous changes in the AFM topology. Only after a subtriangle cordierite-biotite-chlorite has moved toward the F side of the large triangle across the composition of rock O does a new mineral (cordierite) appear. As long as rock O is inside the subtriangle chlorite persists. The net change can be represented as:

chlorite + muscovite + quartz = biotite
+ cordierite + water (R.4)[1]

the continuous reaction proposed for the cordierite isograd in the Slave Province by Davidson (1967) and Kretz (1968) and, with a small modification, by Ramsay (1973a,b,c). Further increase of metamorphic grade produces two discontinuous changes (Fig. 7) in the AFM diagram; but rock O remains in the cordierite-biotite field until the subtriangle cordierite-biotite-andalusite has formed and migrated toward the M side of the main triangle, producing andalusite and/or sillimanite by the reaction:

cordierite + muscovite = andalusite/
sillimanite + biotite (R.6)

[1] Reaction equations are numbered in order of increasing metamorphic grade corresponding to sequence of AFM projections shown in Figure 7.



a) P-T diagram:

light solid lines (1), (3), (5) Hess (1969), (7) Chatterjee and Johannes (1974), (8) Brown and Fyfe (1970);

heavy solid lines, estimates for continuous reactions (4) and (6) for a metagreywacke with M ≃ .6 (Fig. 6);

short-dashed lines, Fe end member (Thompson, A.B. 1976) and Mg end member (Seifert, 1970);

double lines (Richardson et al., 1969);

squiggly lines, melting curves for biotite paragneiss after Winkler (1976);

long-dashed lines, erosion surface P-T curves based on the isograd sequences in (b). Small circles mark the reactions inferred to have occurred at the isograds. For all equilibria except (8) $P_{H_2O} = P_{rock} = P$. The dotted line marked with X's represents the P-T history of the kyanite-staurolite sillimanite-cordierite rock from K_4 (Fig. 1 and 4d).

b) Traverses from the erosion surface in the Slave Province:

A) Point Lake (Bostock, 1976),

B) Sparrow Lake east of Yellowknife (Kamineni, 1973, 1975),

C) Benjamin Lake (Davidson, 1967; Heywood and Davidson, 1969). Circles designate isograds.

Figure 8. Mineral symbols as on Figure 3.

The movement of the cordierite-biotite-chlorite subtriangle toward F and of the cordierite-biotite-andalusite subtriangle toward M with increasing P and T is consistent with the observed isograd sequence and with A.B. Thompson's Table 2 (p. 406-407, 1976).

A rock (X) with a higher concentration of Fe has a different history. The appearance of staurolite when the chlorite-biotite-staurolite subtriangle overrides composition X marks its departure from the chlorite-biotite field (Fig. 7). The staurolite isograd in this case would be represented by the continuous reaction:

$$\text{chlorite + muscovite + quartz = biotite + staurolite + water} \quad (R.2)$$

Subsequently discontinuous reactions (3) and (5) affect rock X:

$$\text{chlorite + staurolite + muscovite = andalusite + biotite + water} \quad (R.3)$$

and

$$\text{chlorite + muscovite + quartz = cordierite + biotite + andalusite + water} \quad (R.5)$$

Figure 7 and the sequence of reactions in rocks of composition X provide an explanation for the occurrences in the Slave Province of staurolite in the cores of andalusite and cordierite grains recorded by Wright (1950), Kamineni (1973), Davidson (1967) and Bostock (1976) (e.g. Fig. 4). Theoretically, in order for staurolite to persist on the high grade side of reaction (3), either chlorite or muscovite or both must be completely consumed, which would prevent the occurrence of reaction (5) in that rock. However, if the distance between (3) and (5) in P-T space is small and/or the rate of metamorphism is sufficiently rapid there is a possibility that staurolite and chlorite would still be in the rock when conditions appropriate to reaction (5) (chlorite out) are attained. In this case, rims of cordierite and andalusite could grow on the staurolite. That is, the staurolite would represent lower grades of the same low pressure metamorphism that produced the cordierite and not an earlier higher pressure metamorphism (Ramsay and Kamineni, 1977). The occurrence of andalusite with staurolite inside cordierite (Fig. 4) supports the conclusion that pressure was not necessarily higher when staurolite formed than it was when cordierite formed.

The P-T Diagram

Construction of the P-T diagram for this study was complicated by insufficient and contradictory experimental data, the variety of petrogenetic grids in the literature, and the correspondence of isograds in the Slave Province with mineral reactions that have not been calibrated experimentally. The diagram in Figure 8a is a reasonable compromise despite several arbitrary features.

The selection of a triple point for the Al_2SiO_5 system and the assumption made about the activity of water (a_{H_2O}) are controversial. The history of the location of the triple point has been dynamic, with estimates ranging from 2.5 to 7 kb* and from 480 to 650°C (Schau, 1978). For most petrologists, the debate now centres on the P-T region between 5.5 kb-622°C (Richardson et al., 1969) and 3.76 kb-501°C (Holdaway, 1971). There is reasonably good agreement on the P-T coordinates of the kyanite-sillimanite and andalusite-kyanite transitions. Holdaway's andalusite-sillimanite transition does not allow for the existence of magmatic andalusite (Schuiling, 1960; Clarke et al., 1976) or andalusite migmatite (Bostock, 1976). Recently, the Richardson-Gilbert-Bell triple point has been applied successfully to the Central Alps (Hänny, 1972; Thompson, 1976), Naxos, Greece (Jansen et al., 1976), the Dalradian (Porteous, 1973; Richardson and Powell, 1976), and

* 1 kb = 10^5 kPa

the Rhodesian craton (Saggerson and Turner, 1976). On the other hand, Ferry (1976) concluded that Holdaway's version applied to his study area in south-central Maine. The widespread occurrence of coexisting aluminosilicates suggests that for field petrologists the triple point is a volume which occurs at different pressures and temperatures in different metamorphic terranes. A major problem is the small energy change for the andalusite-sillimanite transition. With such a small energy change, differential stress, fluid content of the rocks, and grain size may be of major importance. It is conceivable that sillimanite develops at lower temperatures in andalusite-bearing rocks undergoing deformation than in rocks where deformation has ended and that the andalusite-sillimanite transitions of Richardson et al. (1969) and of Holdaway (1971) are the upper and lower limits, respectively, of a range of possible transitions. The one which applies to a particular terrane is only evident from the spatial relations between the stability areas in the field of the polymorphs of Al_2SiO_5 and the rest of the isograd pattern. In the Slave Province, although 5.5 kb-622°C (Richardson et al., 1969) appears to be a little high, this triple point, together with the calculated and experimental data of Chatterjee and Johannes (1974), Brown and Fyfe (1970), Hess (1969), Seifert (1970), A.B. Thompson (1976), and Winkler (1976), results in a P-T diagram that is compatible with the relative stability fields of mineral assemblages observed in the rocks (Fig. 8a).

The calibrated dehydration reactions included on the P-T diagram were determined with $P_{fluid} = P_{rock}$ and $P_{fluid} = P_{H_2O}$ (activity of H_2O = 1). Whether or not the fluid pressure was equal to rock pressure during metamorphism is not known. Also, as some rocks in the Slave Province are graphite- or carbonate-bearing, CH_4 and CO_2 cause P_{H_2O} to be less than P_{fluid} at some localities. In order that a set of P-T values may be associated with the isograds in the Slave Province, fluid pressure is assumed to be equal to rock pressures during metamorphism in the immediate vicinity of isograds that can be represented as fluid-producing mineral reactions. The fluid phase composition is assumed to be controlled by the mineral assemblage (Greenwood, 1975). That is, the presence or absence of a fluid and its composition is taken to be a function of temperature, rock pressure, and rock composition (e.g., activity of H_2O is equal to unity near dehydration reactions). The temperatures inferred from the P-T diagram are, therefore, maximum temperatures.

The locations in Figure 8a of the continuous reactions:

$$\text{chlorite + muscovite + quartz = cordierite + biotite + } H_2O \quad \text{(R.4)}$$

and

$$\text{cordierite + muscovite = biotite + andalusite/sillimanite} \quad \text{(R.6)}$$

are crude estimates that apply only to one composition, a typical Slave metagreywacke (M ~ 0.6, Fig. 6). The reactions must intersect at the discontinuous reaction:

$$\text{chlorite + muscovite + quartz = andalusite + cordierite + biotite + } H_2O \quad \text{(R.5)}$$

(A.B. Thompson, 1976) with reaction (4) on the high temperature side of reaction (5). The curve along which 50 per cent of biotite paragneiss has melted in the presence of enough water to saturate the melt is extrapolated from Winkler (1976). This curve is an approximation in P-T space of the migmatite isograd in the Yellowknife Supergroup.

It is important to emphasize that the method used in this paper to determine the distribution of metamorphic temperatures and pressures from isograd patterns is independent of the P-T diagram used. When an accurately calibrated petrogenetic grid is available, the exercise carried out here can be redone. The absolute temperatures and pressures associated with an isograd may change but the relative positions of mineral assemblage stability fields are not likely to differ significantly from those in Figure 8a. Furthermore, the relationship between erosion surface P-T curves and geothermal gradients discussed in the next section will remain the same.

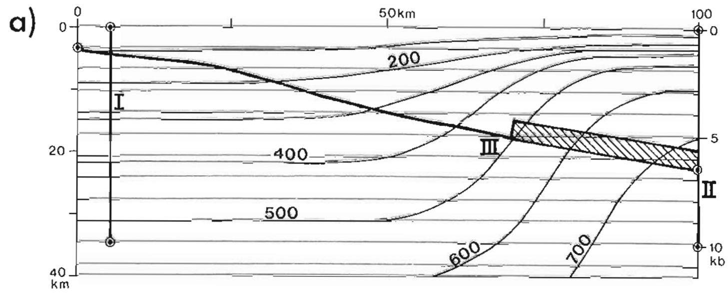


Figure 9a. P-T profile through part of the distribution of temperature and pressure at the peak of Lepontine metamorphism (Swiss Alps) (after Thompson, 1976, 1977). I and II are geothermal gradients. III is an estimate of the present day erosion surface.

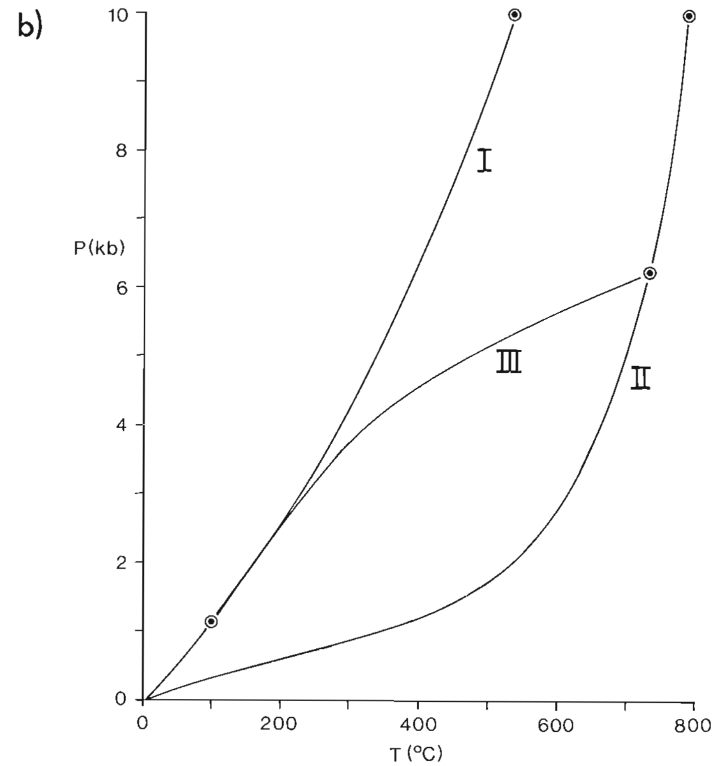


Figure 9b. P-T diagram illustrating the relationship between geothermal gradients (I and II) and the erosion surface P-T curve (III). Isotherm values are degrees Celsius.

Erosion Surface P-T Curves and Geothermal Gradients

Traverses across the erosion surface in the Slave Province provide different isograd sequences, both with respect to the mineralogical changes at the isograd and to their spacing (Fig. 8b). The traverses are based on the work of Davidson (1967), Kamineni (1972), and Bostock (1976). Correlation of the isograds with mineral equilibria on the P-T

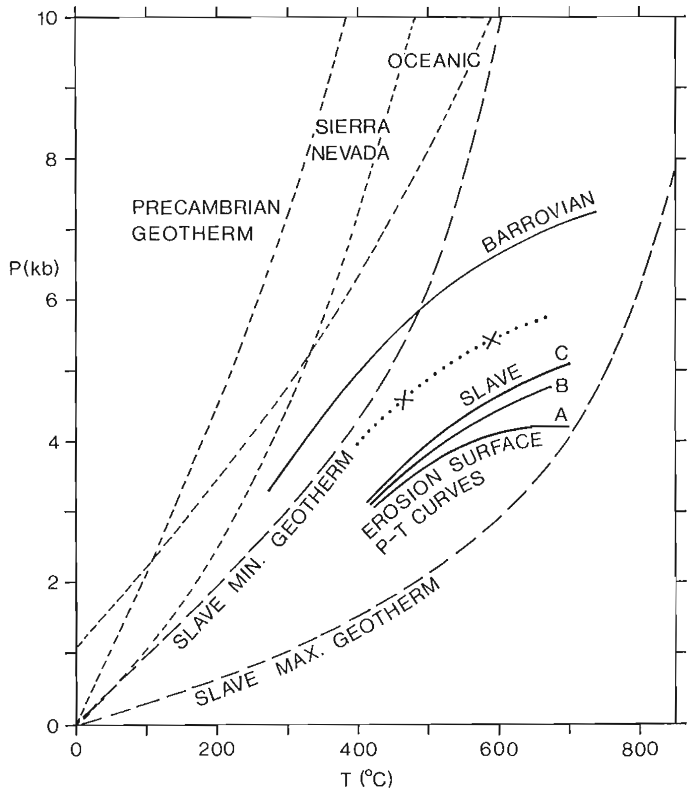


Figure 10. P-T diagram with geothermal gradients calculated by geophysicists (Precambrian and oceanic after Ringwood, 1966; Sierra Nevada after Lachenbruch, 1968), the maximum and minimum geothermal gradients for the peak of metamorphism in the Slave Province, erosion surface P-T curves from the Slave Province, and an erosion surface P-T curve for Barrovian metamorphism (after Turner, 1968). The curve marked by X's represents the P-T history of the kyanite-staurolite-sillimanite-cordierite rock (K_4 Fig. 1 and 4d).

diagram permits the plotting of the relation between pressure and temperature on the erosion surface as an erosion surface P-T curve (Fig. 8a). The P-T curves obtained from the Slave Province erosion surface traverses have the same form as those derived by Hietanen (1967) and Turner (1968) as a basis for distinguishing metamorphic terranes; that is, both pressure and temperature increase with metamorphic grade.

The rarity of staurolite coexisting with cordierite and muscovite and the common occurrence of andalusite with biotite suggest pressures greater than Hess's (1969) invariant point (a, Fig. 8a) at all three localities. The different isograd sequences above the cordierite isograd can be explained by differences in pressure. Whereas the coexistence of andalusite with K feldspar at Point Lake (Bostock, 1976) implies pressures close to the intersection of reaction (7) and the andalusite/sillimanite transition, the isograd pattern at Benjamin Lake (Davidson, 1967) requires a higher pressure P-T curve. Furthermore, the rocks with cordierite enclosing staurolite and kyanite from the eastern Slave Province must have been in the stability field of kyanite during their metamorphic history. A P-T-time curve for these rocks (X...X) is included in Figure 8a. The coincidence of the first appearance of cordierite and the first appearance of andalusite near Point Lake (J.B. Henderson, pers. comm., 1977) probably reflects a slightly higher average Fe content which permits reaction (4) to proceed in the Point Lake rocks. Although data are limited, there is support for the conclusion that more than one erosion surface P-T curve can be derived from the Slave Province.

Geothermal gradients, the variation of temperature with depth beneath the earth's surface, can be plotted on a P-T diagram if a density for the rocks is assumed. Where isotherms parallel isobars the temperature is fixed at a given pressure during metamorphism and there will be only one geothermal gradient across the entire terrane; in this case only, the erosion surface P-T curve corresponds to the curve representing the geothermal gradient. Also, regardless of the orientation of the present erosion surface with respect to the earth's surface during metamorphism, there will be only one erosion surface P-T curve (Thompson, 1977). If isotherms are oblique to isobars (e.g. thermal dome, Fig. 9a) the geothermal gradient varies across the dome (Richardson, 1970) and a range of erosion surface P-T curves is possible. All the possibilities must lie between the limiting geothermal gradients (Fig. 9b). The erosion surface P-T curves differ from the geothermal gradient curves in that the change of pressure along the erosion surface is much less than the change in pressure along the geothermal gradient (i.e. vertically). The decrease of the geothermal gradients derived from the profile (Fig. 9) with increasing temperature (isotherm spacing increases with increasing temperature) is consistent with geothermal gradients calculated by geophysicists on the basis of present surface heat flow, seismic data, and an assumed distribution of heat sources (e.g. Lachenbruch, 1968).

To proceed further, the relationship between the erosion surface P-T curve and geothermal gradient curves (Fig. 9) is applied to Archean metamorphism in the Slave Province. The range of erosion surface P-T curves derived above (Fig. 8a) for this province requires that isotherms be oblique to isobars and that geothermal gradients vary between a maximum and a minimum gradient. Two geothermal gradients which include all the variation presently known in the Slave Province were estimated (Fig. 10). An erosion surface P-T curve from Barrovian terrane and geothermal gradients calculated by geophysicists are included for comparison. The maximum and minimum gradients were used to construct the P-T profile described in the next section.

P-T Profiles and the Distribution of Pressure and Temperature

Before regional metamorphism can be integrated into the geological history of a terrane like the Slave Province, some understanding of the distribution of pressure and temperature in the rocks during metamorphism is required. The isograd pattern records the variation of temperature and pressure in the rocks when the isograds formed. The distribution of pressure and temperature in a portion of the crust can be represented by the temperature gradient (°C/km) perpendicular to isotherms, the pressure gradient (kb/km) perpendicular to isobars, and the angle (α) between the gradients (Thompson, 1976). Temperature gradient as used here is not the same as geothermal gradient; wherever isotherms are oblique to isobars ($\alpha \neq 0°$) the temperature gradient is oblique to the geothermal gradient. The distribution of temperature and pressure in the crust can be represented by a P-T profile, a diagram which illustrates the relationships between temperature, pressure, and distance (vertical and horizontal) (Fig. 11). The P-T profile may be schematic or, if the vertical dimension is exposed, a true cross-section. In most cases, a profile is constructed for a particular time, e.g., when an isograd pattern formed.

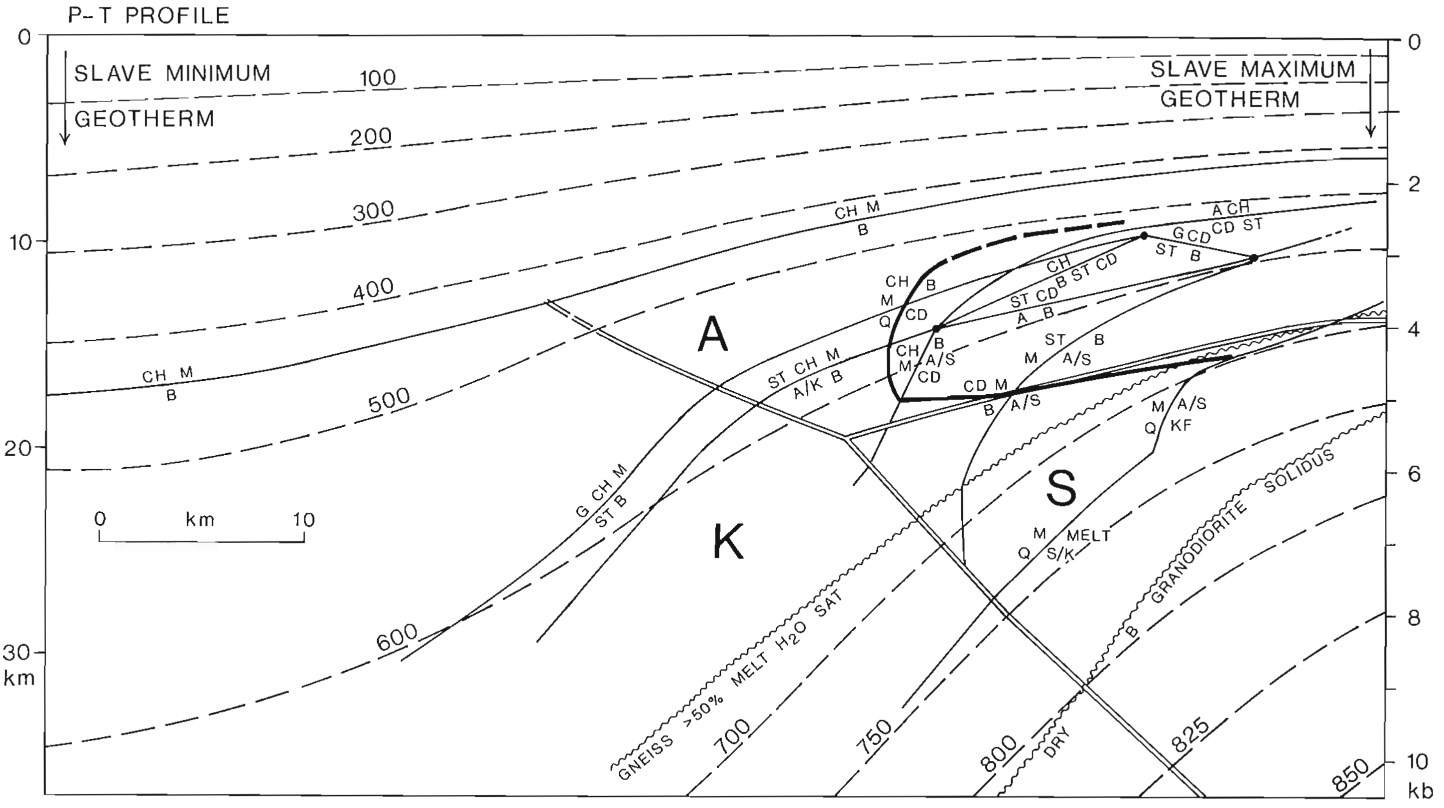


Figure 11a. P-T profile of a hypothetical metamorphic P-T distribution in the earth's crust. Mineral abbreviations as in Figure 3. Mineral reactions plotted include parts of Hess's (1969) petrogenetic grid, the continuous reactions (4) and (6) from Figure 8, quartz-muscovite-stability after Chatterjee and Johannes (1974), the biotite-producing reactions and melting curve (extrapolated) of Winkler (1976), and the solidus for dry biotite granodiorite (Brown and Fyfe, 1970).

Mineral equilibria may be plotted on a P-T profile just as they are on a P-T diagram; but distance, an important variable on a profile, cannot be represented accurately on the P-T diagram. Furthermore, in a volume of rock undergoing metamorphism the spacing of isotherms and their angular relation to isobars is variable and, consequently, so is the geometry of the isograd pattern. For these reasons, the geometry of the isograd pattern cannot be related directly to the pattern of equilibria on a P-T diagram.

Using the minimum and maximum geothermal gradients estimated for the Slave metamorphism (Fig. 10) and an assumed pressure gradient of 0.286 kb/km (3.5 km ≡ 1 kb) a P-T profile has been constructed that accounts for the variations in metamorphic grade observed in the Slave Province (Fig. 11a). A thermal dome in a regional metamorphic terrane is not a new idea; for example, domes have been proposed for the Dalradian (Kennedy, 1948), the Alps (Wenk, 1970), and the Benjamin Lake area of the Slave Province (Davidson, 1967). It is necessary, however, to differentiate the real thermal domes, isotherms oblique to isobars ($\alpha > 0^o$), from tectonic thermal domes (e.g. Niggli, 1970), the structural doming of originally horizontal isotherms, isobars, and isograds ($\alpha = 0^o$). The range of erosion surface P-T curves (Fig. 10) and geothermal gradients in the Slave Province is possible only if isotherms were oblique to isobars (Fig. 11a) when the isograds formed. Erosion surfaces cutting into one dome at different levels or into several domes of different configurations will produce the variety of erosion surface P-T curves obtained. Differential erosion of an irregularly spaced group of thermal highs and lows similar to the one in Figure 11a will produce the patchy pattern of low and high grade zones separated by an extensive medium grade zone preserved in the Slave Province (Fig. 1). Although there is a fairly regular 200 km spacing of the major low grade areas, the pattern at Yellowknife, Point Lake, and in the southeastern corner of the province indicates that the spacing is variable, and may be less than 30 km. It is interesting that as a thermal dome of this configuration expands laterally through crust rocks originally in the kyanite-staurolite field (Fig. 11a), the rocks can enter the cordierite-sillimanite field; no decrease in pressure is required. The cordierite-rimming-kyanite texture previously mentioned (Fig. 4) can be explained this way.

To illustrate the effect of a thermal high on the various rock types in the Slave Province, a P-T profile has been superimposed on a schematic section through an isoclinally folded sequence of greywacke and basalt (Yellowknife Supergroup) and biotite granodiorite (basement) (Fig. 11b). At the transition from low to medium grade the cordierite and staurolite isograds correspond closely to the transition from metabasalt to amphibolite, but no major changes occur in the basement rocks. At higher grades in the core of the dome, complex melting relationships are possible with metasediments and basement melting at different pressures and temperatures; the amphibolite is not affected. The profile illustrates how granitoid rocks produced in the core of a thermal dome may intrude overlying metamorphic rocks related to the same thermal dome, a relationship commonly observed in the Slave Province.

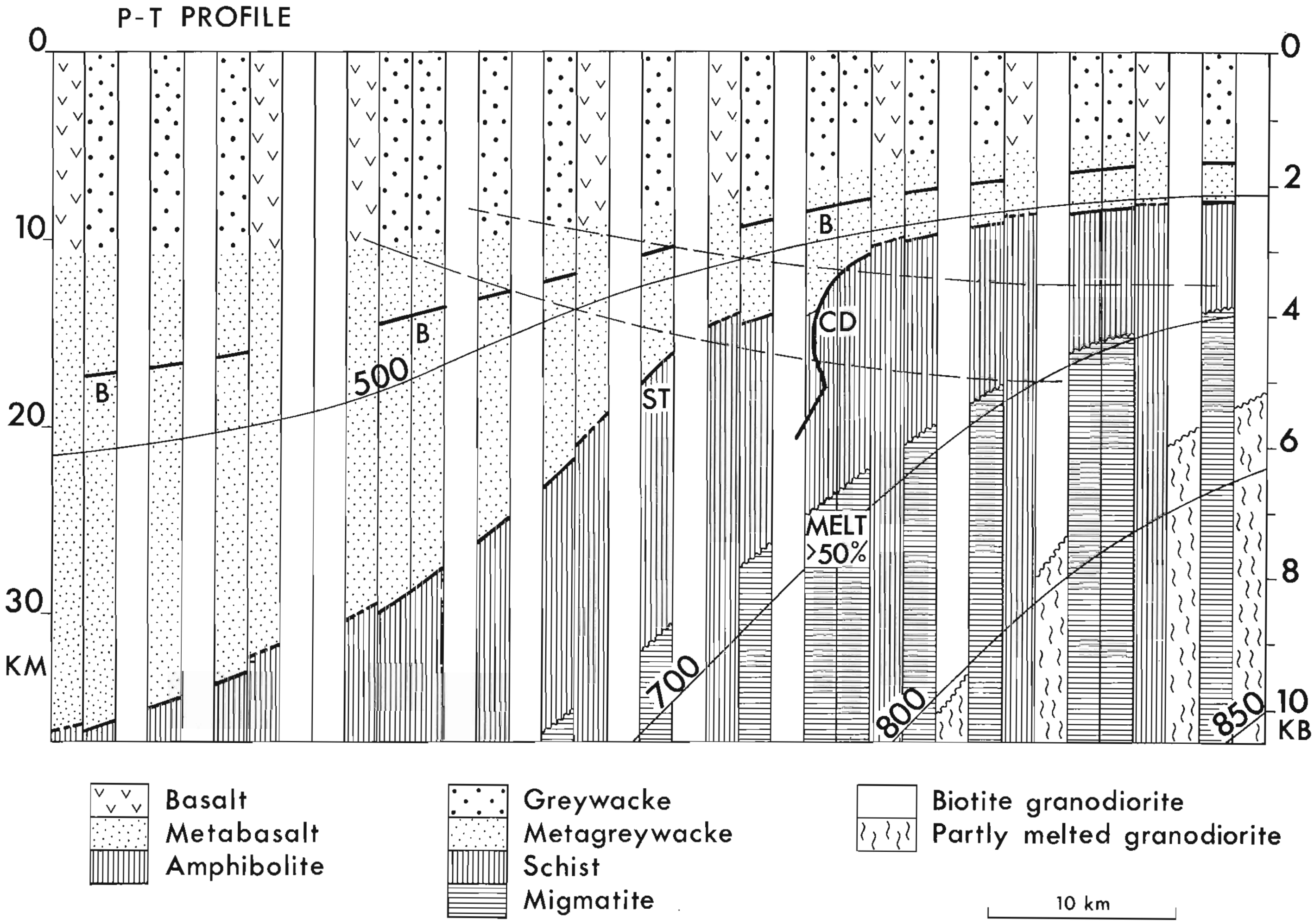


Figure 11b. P-T profile from (11a) superimposed on schematic crust composed of rocks similar to those in the Slave Province. B, CD, ST, MELT, designate isograds and the dashed curves possible erosion surfaces. Isotherm values are degrees Celsius.

CRUSTAL RECONSTRUCTION

Accepting the presence of sialic basement beneath the Yellowknife Supergroup, the erosion surface P-T curves and geothermal gradients (Fig. 10) can be used to construct a cross-section (Point Lake-Lac de Gras) of the crust as it is now and to reconstruct another as it was 2600 Ma ago at the peak of metamorphism (Fig. 12). Whereas the isograd dip is constrained to some extent by the limiting geothermal gradients and the low temperature at the earth's surface, the basement/cover contact farther from the present erosion surface is highly speculative. The present crustal thickness (35 km) is taken from Barr's (1971) data for the southern part of the province. Assuming an average rock density of 2.85 (3.5 km ≡ 1 kb), the erosion surface P-T curves (Figs. 8a, 9) imply postmetamorphism uplift and erosion ranging from less than 5 km in low grade areas to more than 15 km in areas of high grade. The thickness of the present crustal section combined with the eroded section, implies a crust 40-50 km thick 2600 Ma ago. This number is much greater than that proposed for Archean terranes by Glikson (1972), Anhaeusser (1969), or Hargraves (1976), but compatible with data from the Archean of Scotland (Dickinson and Watson, 1976) and Greenland (Wells, 1976). The superposition of variable but locally high geothermal gradients attributed to the metamorphism in the Slave Province on a crust of this thickness could produce, by partial melting of pre-existing sialic material, most of the large proportion of granitoid rocks presently exposed.

The hypothesis that the formation and the thickening of the Archean crust occurred by addition of granitic material derived from the mantle and that deformation/metamorphism of supracrustal rocks was caused by the intrusion of the granitoid rocks has been applied to the Slave Province (Green and Baadsgaard, 1971) as well as to other Archean terranes (e.g., Glikson, 1972; Anhaeusser, 1969). In the Slave Province, the presence of granitic rocks, basement to supracrustal rocks, and the fact that many plutonic masses postdate a regional metamorphism which attained maximum grade after the main phases of deformation had ceased, suggest that an alternative hypothesis merits consideration. For example, high geothermal gradients in areas of sialic crust thickened by horizontal shortening could have resulted in the metamorphism of sedimentary and volcanic rocks and, at depth, the partial melting of these rocks and underlying sialic basement, culminating in the intrusion of the granitoid rocks. Crustal shortening (thickening) may have occurred in Archean rocks of Greenland (Bridgwater et al., 1974) and the Rhodesian craton (Coward, 1976). Horizontal compression was tentatively proposed (Fyson, 1975) to account for the prominent schistosity in rocks east of Yellowknife (Fig. 1).

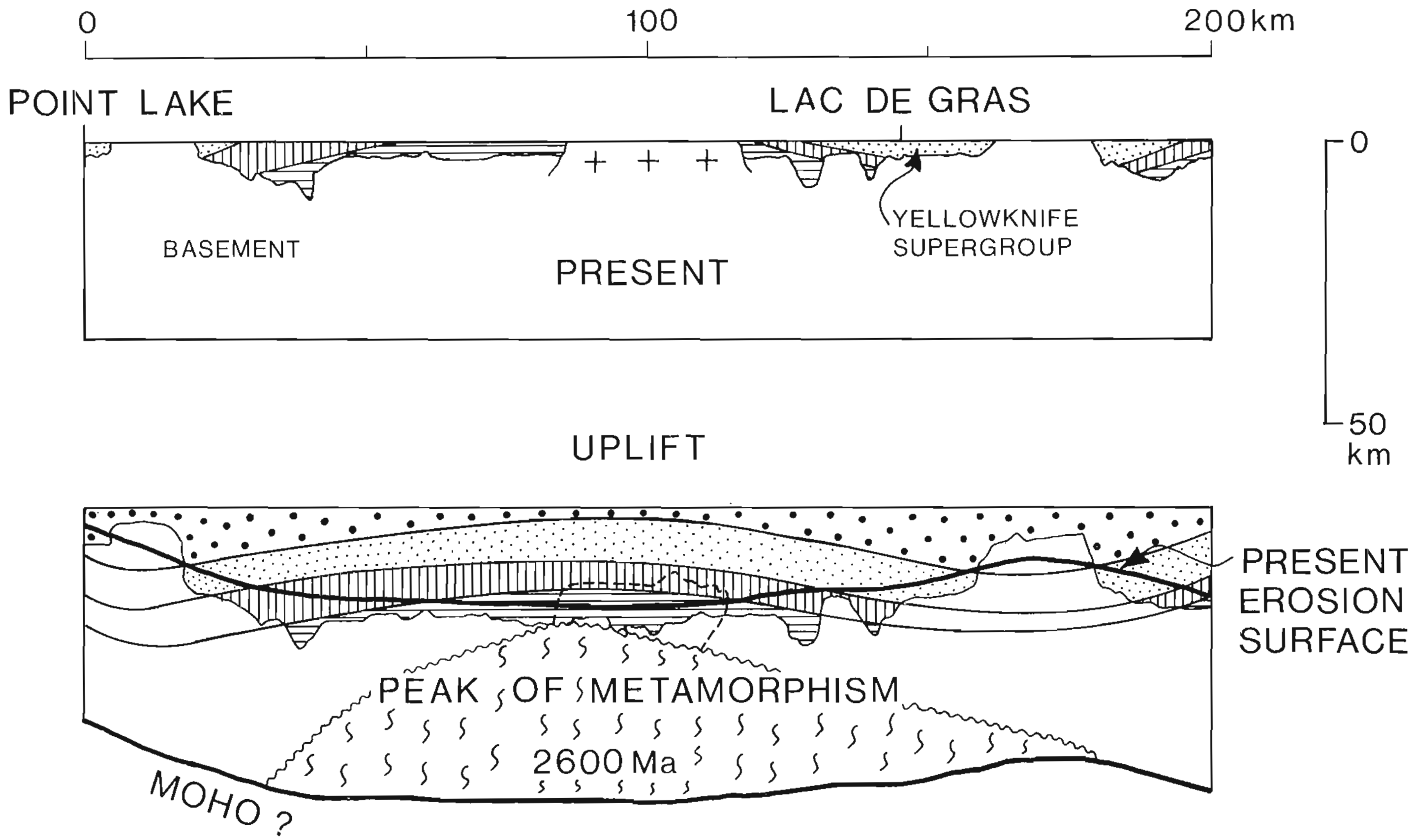


Figure 12. Reconstruction of the crust between Point Lake and Lac de Gras (Fig. 1) at present and at the peak of metamorphism about 2600 Ma ago. Supracrustal rocks are patterned: widely spaced dots – unmetamorphosed; closely spaced dots – low grade; vertical lines – medium grade; horizontal lines – high grade; squiggles – partly melted sialic basement rocks.

Production of granitoid rocks from deep-seated sialic crust by a thermal dome that metamorphosed supracrustal rocks nearer the surface has been invoked for the Benjamin Lake area in the Slave Province (Davidson, 1967), the Rhodesian craton (Saggerson and Turner, 1972, 1976), and the Yilgarn Block (Archean) in Australia (Binns et al., 1976). The crust 2600 Ma ago in the Slave Province, e.g., the Point Lake-Lac de Gras rocks as reconstructed in Figure 12, may have been produced by a combination of tectonic thickening caused by horizontal compression and geothermal gradients ranging between the limiting gradients derived in this paper. Testing of this alternative working hypothesis will require that metamorphism and deformation be emphasized in future mapping in the Slave Province.

CONCLUSIONS

The patchy distribution of low pressure regional metamorphism in the Slave Province is the result of differential erosion of an irregular thermal topography which attained its maximum development after the major phases of deformation had ceased. Locally high geothermal gradients imposed on a 40-50 km section of sialic crust produced the low pressure metamorphism and, at depth, caused widespread partial melting culminating in the intrusion of most of the granitoid rocks presently exposed.

For some rock compositions continuous mineral reactions involving the migration of a three-phase subtriangle across the face of a phase diagram represent isograds more accurately than discontinuous reactions involving an exchange of tielines.

In a two-dimensional metamorphic terrane, mineral assemblages and the variable spatial relations of isograds can be used to reconstruct the three-dimensional distribution of pressure and temperature in the crust when the isograds formed. Until a generally acceptable petrogenetic grid is available, it is worthwhile to reconstruct metamorphic P-T distributions on the basis of available experimental data. A better understanding of the field relations in metamorphic rocks imposes important constraints on the ideal grid.

The attempt made in this paper to incorporate a synthesis of regional metamorphism into the Archean tectonic history of the Slave Province presents some interesting problems. The extent to which sialic crust was thickened by horizontal compression during the 2600 Ma tectonic event, the proportion of high grade rocks formed by a pre-2600 Ma metamorphism, and the timing of the development of the irregular thermal topography across the province are intriguing questions for the future.

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TECTONICS AND METAMORPHISM ALONG THE SOUTHERN BOUNDARY BETWEEN THE BEAR AND SLAVE STRUCTURAL PROVINCES

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Frith, R.A., Tectonics and metamorphism along the southern boundary between the Bear and Slave structural provinces; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 103-114, 1978.

Abstract

Recent structural, mineralogical, and geochronological data from the Bear-Slave boundary region have permitted subdivision of Archean and Proterozoic orogenic events into tectonic episodes, each with associated thermal events.

The Yellowknife Supergroup volcanic and sedimentary rocks of the Slave Province have been deformed during two largely compressional Archean episodes and one largely dilational Proterozoic episode. The Archean episode consisted of three phases. An initial open folding (F_1) was accompanied by chlorite grade metamorphism (M_1). A second deformation (D_2) tightly folded the metasediments into isoclines and vertically stretched the volcanic rocks contemporaneously with the development of a biotite schistosity and local development of porphyroblasts. Overgrowths of earlier porphyroblasts and the development of new porphyroblasts followed, as part of the same long-lived thermal phase (M_2). Granitic intrusions postdate D_2 and M_2 and are associated with low grade regional metamorphism (M_3) and local brittle deformation. The Proterozoic dilational deformation was reached by differential vertical movement. Isotopic systems record two thermal events, both associated with minor intrusive activity, at about 2200 Ma (M_4) and 1900 Ma (M_5) involving the intrusion of diabase dykes, granodiorite stocks and, the formation of granitic pegmatite.

The Bear-Slave boundary principally separates Archean cratonic rocks from exhumed Archean basement and Proterozoic cover rocks (Snare Group). Where the cover is thin, aeromagnetic patterns transgress the boundary and isotopic data reflect early Proterozoic pre-Snare thermal episodes found within both the Slave and Bear provinces. The cover rocks range in grade from greenschist facies along the boundary to amphibolite facies. The granulite facies terranes are associated with diapiric uplifts of granitic and supracrustal rock.

Résumé

Les données récemment obtenues sur la structure, la minéralogie et la géochronologie de la limite entre la province de l'Ours et la province des Esclaves ont permis de subdiviser les cycles orogéniques de l'Archéozoïque et du Protérozoïque en plusieurs phases tectoniques, qui ont toutes été accompagnées de phénomènes thermiques.

Les roches volcaniques et sédimentaires du supergroupe de Yellowknife, dans la province des Esclaves, ont été déformées au cours de deux épisodes de compression pendant l'Archéozoïque et d'un épisode, de dilatation principalement, pendant le Protérozoïque. L'épisode protérozoïque se subdivise en trois phases. Tout d'abord, une phase de plissement initial ouvert (F_1) a été accompagnée du métamorphisme caractéristique de la zone des chlorites (M_1). Une seconde déformation (D_2) a fortement plissé les métasédiments en plis isoclinaux, et verticalement étiré les roches volcaniques; en même temps, il s'est formé des schistes à biotite, et en certains endroits, des porphyroblastes. Les porphyroblastes existants ont continué à s'accroître, de nouveaux porphyroblastes se sont formés pendant la même phase thermique prolongée (M_2). Les intrusions granitiques sont ultérieures à D_2 et M_2, et ont donné lieu à un métamorphisme régional d'intensité faible (M_3), et localement à des phénomènes de déformation et fracturation. Les déformations par dilation qui ont eu lieu au Protérozoïque résultaient de mouvements différentiels verticaux. Deux évènements thermiques, que l'on a datés à environ 2 200 Ma (M_4) et 1 900 Ma (M_5) ont laissé leur marque; il y a en particulier eu intrusion de dykes de diabase, mise en place de stocks granodioritiques, et formation de pegmatite granitique.

La limite entre la province de l'Ours et la province des Esclaves sépare principalement les roches cratoniques qui émergent du soubassement archéozoïque, et les roches de couverture d'âge Protérozoïque (groupe de Snare). Là où la couverture est mince, les structures décelées par les levés aéromagnétiques débordent la limite, et les données isotopiques indiquent l'existence d'épisodes thermiques d'âge protérozoïque antérieurs au groupe de Snare, dans la province des Esclaves et la province de l'Ours. Les roches de couverture vont du faciès schistes verts le long de la limite entre les deux provinces, au faciès amphibolite. Les terrains métamorphisés dans le faciès granulite sont associés à des soulèvements diapiriques des roches granitiques et supracrustales.

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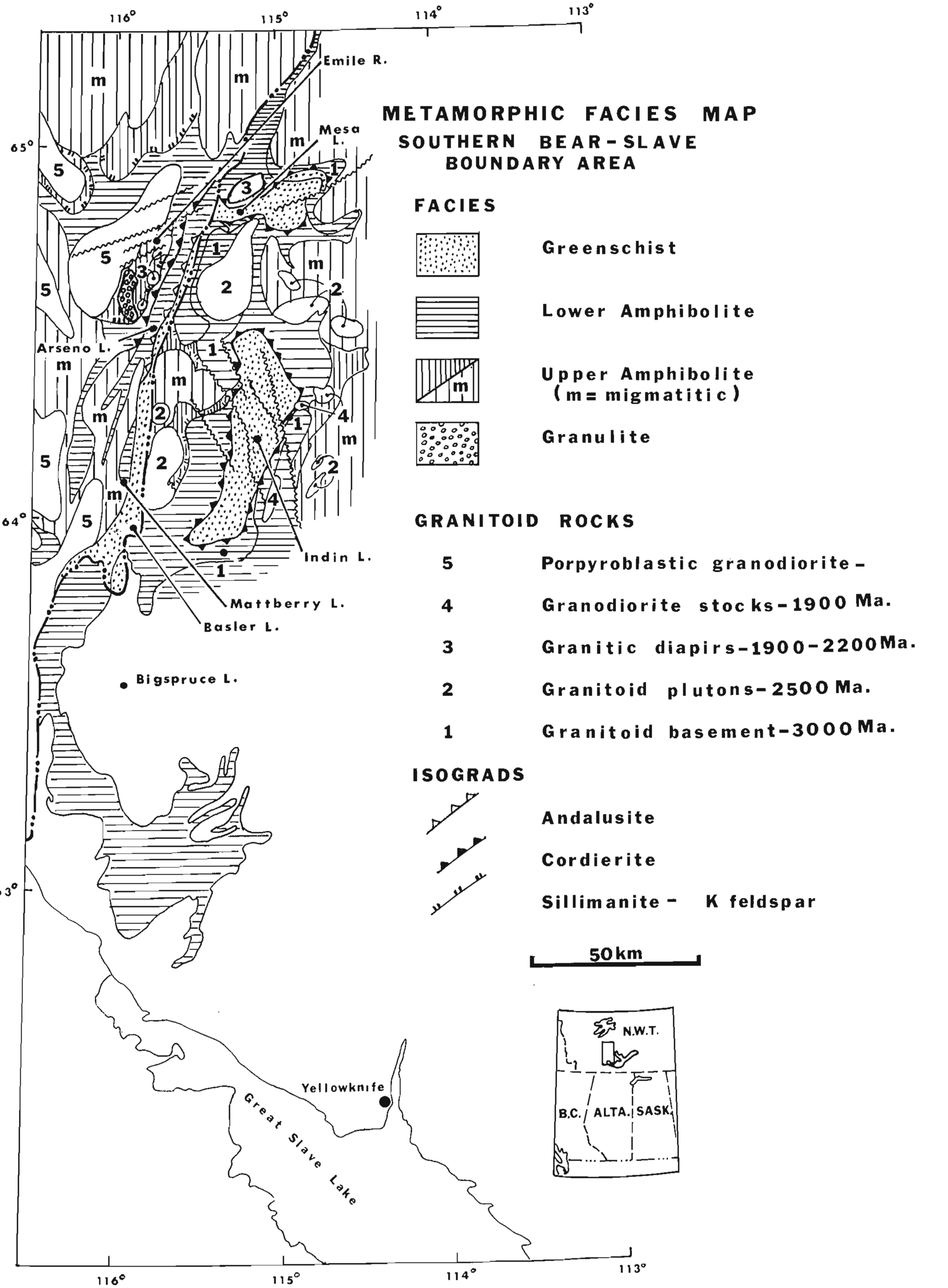

Figure 1. Metamorphic facies map of the southern Bear-Slave boundary area showing the distribution of isograds and the approximate age of granitoid rocks in the region. Where no isograd exists between facies represented, age discrepancies or dislocations are implied.

INTRODUCTION

The metamorphic history of the southwestern Slave Province (Fig. 1) is related to tectonism of both the "Kenoran" and "Hudsonian" orogenies. Altogether, six thermal episodes are recognized, three during the Archean and three during the Proterozoic. Because the effect of these thermal episodes is variable from place to place, each thermal episode can be related sequentially to the deformational history of the region. Radiometric dating is sufficiently advanced that the metamorphism can be placed in a chronometric framework.

Advances in structural geology, metamorphism, and geochronology have resulted from recent mapping at scales of 1:250 000 and 1:50 000 and laboratory studies (Frith, 1973; Rosaline Frith, 1973; Frith et al., 1974; Frith and Leatherbarrow, 1975; Frith et al., 1977). This work builds on previous studies of the Bear-Slave boundary region (McGlynn and Ross, 1961; Ross, 1959, 1966; Ross and McGlynn, 1965; Smith, 1966). Regional mapping of the area was carried out by J. Tuzo Wilson and C.S. Lord (Lord, 1942) and by Fortier (1949). More detailed mapping of the Indin greenstone belt of the region was carried out by Tremblay (1948), Stanton (1947), and Wright (1950).

GENERAL GEOLOGY

The southwestern Slave Province is made up of five principal rock groups:

(1) A granitoid basement consisting mainly of tonalitic to trondhjemitic gneiss with some granite and granodiorite is about 3000 Ma* old. The rocks are commonly migmatized or metasomatized and intruded by an east-west trending set of gabbro dykes which are locally truncated by volcanic belts about 2700 Ma old. Rocks near the volcanic belts may be extensively mylonitized.

(2) Yellowknife Supergroup volcanic rocks which presumably overlie the basement are about 2700 Ma old. Pillow tops face away from the basement-volcanic contacts. The volcanic belts are variable in thickness and distribution, due principally to the effects of tectonism. On a macroscopic scale the belts, as much as 10 000 m thick, present a pinch and swell appearance. They are made up mainly of pillowed basalts capped with about 10 per cent (by area) rhyolitic breccias and tuffs.

(3) Yellowknife Supergroup metasediments, about 2700 Ma old, for the most part overlie the volcanic rocks. Locally, thin sequences of volcanic rocks are interfingered with the sediments. At the contact, volcanic pebble conglomerate and volcanic-granitoid pebble paraconglomerate may be present. The true thickness of the metasedimentary pile is unknown, due in part to tight isoclinal folding.

(4) Granitic and migmatitic rocks, about 2500 Ma old, intrude as plutons, as migmatitic-supracrustal-plutonic complexes, and as remobilized 2700 or 3000 Ma old granitic-migmatitic complexes. The emplacement of these rocks was the last Archean event to have taken place. Most contain a fabric that can be related to their uplift, particularly at their margins.

(5) Minor intrusive diabase dykes, granodioritic stocks and granitic pegmatites, about 1960 Ma old, make up less than 5 per cent of the total surface exposures. These rocks were formed as a result of Proterozoic tectonic events related in time, if not origin, to activity within the Coronation Geosyncline.

Deformation in the Slave craton is three-fold. Two major compressional events occurred in the Archean and one dilational event in the Proterozoic. The earliest folding trends east-northeast. Basement rocks, particularly where in contact with east-trending volcanic rocks, are mylonitized, whereas the volcanic rocks are stretched along steep, near vertical, axes and locally converted to banded hornblende gneiss. The metasedimentary rocks commonly have been folded and refolded into tight isoclinal structures (Fig. 2). The complexity varies generally with increase in metamorphic grade. In high grade areas only the second phase of folding is evident. A third minor brittle deformation is widespread but erratic in distribution.

The southeastern Bear Province was penetratively deformed during Proterozoic orogeny in the Coronation Geosyncline. Both Archean basement and Proterozoic cover rocks were affected; basement rocks were remobilized and emplaced into the cover rocks as gneiss domes and large scale intrusions, and the cover rocks were folded during one or more phases.

The basement rocks comprise mostly granitic gneiss that has been migmatized or metasomatized. Large K feldspar porphyroblasts, locally deformed into augen, are present in some of the larger bodies that show rapakivi textures. Other exposures of granitic basement are similar in appearance to Slave craton rocks. Metasedimentary equivalents of the Yellowknife Supergroup that occur within the Bear Province have undergone three or more phases of intense deformation (Fig. 3), and thus contrast with the overlying Snare Group Proterozoic rocks adjacent to the Bear-Slave boundary which have been deformed only once. The simple deformational style of the cover rocks becomes more complex toward the west where the regional metamorphism climbs from lower greenschist to upper amphibolite facies.

METAMORPHISM AND TECTONISM IN THE SLAVE PROVINCE

The southwestern border region of the Slave Province was affected by a pre-Kenoran event, the "Kenoran" Orogeny.

Two relatively lesser known Proterozoic events that culminated in the "Hudsonian" Orogeny are more evident in the Bear than in the Slave Province.

Pre-Kenoran metamorphism

Rb-Sr studies of tonalitic gneiss to the east and northeast of the Indin Lake (greenstone) Belt (Frith et al., 1974) have indicated the presence of rocks 2940 Ma in age. Field studies show that this Rb-Sr age may date an episode of potash metasomatism. Zircons from the same rocks suggest that the age may be 3100 Ma (Wanless, pers. comm., 1976). This isotopic work corroborates stratigraphic relationships that indicate that the gneiss and the east-west dykes that cut them are older than the overlying Yellowknife Supergroup volcanic rocks. Other examples of similar stratigraphic relationships have been reported from the Slave Province (Frith et al., 1977).

The deformation of the pre-Kenoran gneiss can not be reliably dated. Only evidence of the most recent, presumably Kenoran, deformation is left intact. However, cobbles of granitic gneiss at the base of the volcanic pile at Brislane Lake in the southern part of the Slave Province contain a gneissic fabric that predates the fabric in the overlying volcanic sequence (A. Davidson, pers. comm., 1977). These age and deformational data suggest that a pre-Kenoran metamorphic event may have taken place.

* Age data reported in this paper use constants reported by Jaffey et al. (1971) for U-Pb; Beckinsale and Gale (1969) for K-Ar, and Neumann and Huster (1976) for Rb-Sr.

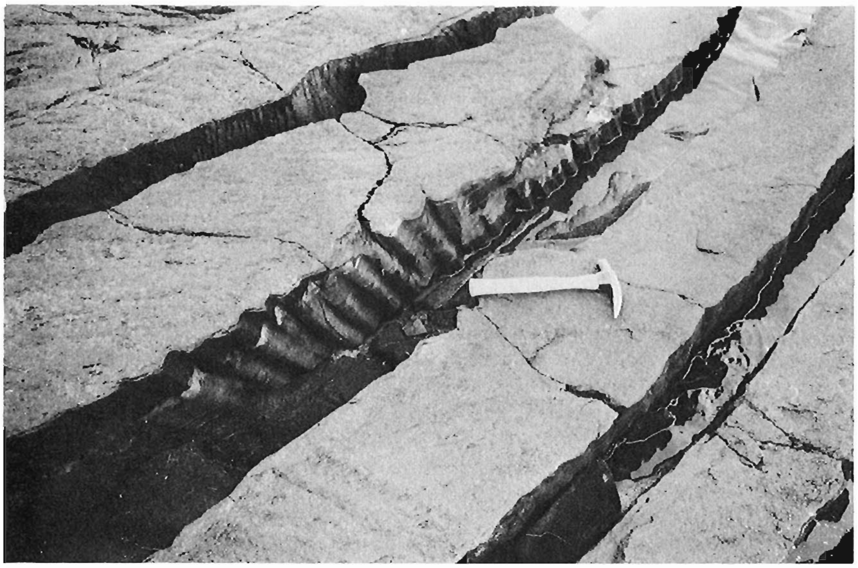

Figure 2

The Yellowknife Supergroup greywackes are doubly folded over much of the area. The first metamorphism (M_1) to affect these rocks was from chlorite to biotite grade and parallels the bedding (argillite is preferentially weathered out). A second more intense metamorphism (M_2) is accompanied by the development of a muscovite schistosity (axes parallel to hammer handle) containing post-schistosity cordierite, sillimanite, and two phases of andalusite.

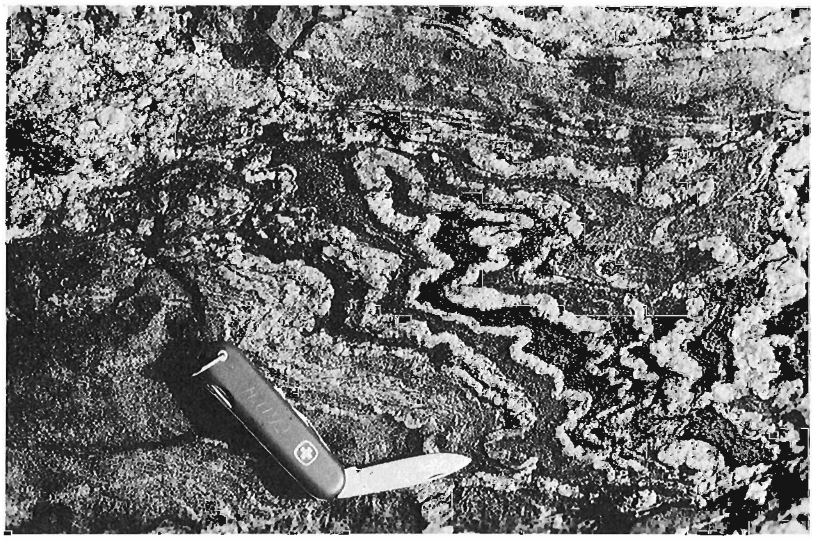

Figure 3

Deformed "Yellowknife" metasedimentary rocks from the exhumed Archean basement in the Bear Province. Proterozoic cover rocks are deformed only once with axes parallel to the last phase of folding shown here (parallel to knife blade). All metamorphic porphyroblasts are completely retrograded to a chloritic matt.

Kenoran metamorphism

The "Kenoran" Orogeny in the Slave Province is analagous to the Kenoran Orogeny of the Superior Province (Stockwell, 1961, 1972). The supracrustal lavas and sediments and the syndepositional igneous intrusions have been deformed and metamorphosed. The depositional age based on zircon ages (Green, 1968; Wanless, 1970) is presumed to be between 2700 and 2600 Ma. However, the volcanic Rb-Sr isochrons, presumably equilibrated by thermal events, took place at 2570 Ma; the isochrons have initial ratios up to 0.706. Granitic rocks at Prosperous Lake which postdate both the deformation and regional metamorphism were, according to the Rb-Sr isochron, reset about 2520 Ma ago (Green, 1968). In the southwestern Slave Province, the large scale granitic intrusions are 2517, 2475, and 2507 Ma in age (Frith et al., 1977) but initial $^{87}Rb/^{87}Sr$ ratios suggest that some are intrusive and some were formed during a period of regional metamorphism.

The Indin Lake supracrustal rocks presently occupy a faulted structural basin similar in outline to that enclosed by the isograd in Figure 1. Low greenschist facies rocks are located in the centre of the basin and the rocks rise in grade both to the east and west. The west side of the basin is intruded by batholithic and migmatitic complexes, whereas the east and south margins are faulted against pre-Kenoran basement gneiss. The contact is intruded locally by Proterozoic granodiorite stocks and granitic pegmatites. The rocks in the northern margin of the structural basin form a south-plunging synformal structure.

The rocks may be generally divided into three metamorphic categories that have been called by earlier workers "cold sediments", "spotted schists", and "nodular schists".

"Cold Sediments"

These sediments are principally greywacke, phyllite, and slate and commonly contain minerals such as sericite, chlorite and biotite, characteristic of low greenschist facies metamorphism. Field classification is based on the degree to which biotite is developed and although "cold sediments" appear to lack biotite, the mineral can commonly be found in thin section. The greywacke contains fewer phyllitic minerals and more quartz and plagioclase. K feldspar was not observed. The phyllitic minerals developed parallel to F_1 axial planes. These surfaces may have been refolded during a second period of folding (F_2). The metamorphism that produced the phyllitic minerals (M_1) is probably contemporaneous with F_1 folding.

"Spotted Schists"

Higher grade metamorphism has destroyed some of the heterogeneity of the lower grade rocks and imparted a more uniform schistose appearance. The finer grained more argillaceous metasediments contain visible spots of biotite in a groundmass of biotite, sericite, and chlorite. The coarser grained varieties appear relatively unchanged from the "cold sediments", as they lack the necessary components for porphyroblast development. The biotite porphyroblasts in the more argillaceous rocks are developed both parallel and transverse to F_1 fabric, suggesting that they continued to grow after F_1.

"Nodular schists"

The "nodular schists" are coarser grained and more homogeneous than their lower grade equivalents. Where the composition was appropriate, porphyroblasts of one or more of the minerals cordierite, andalusite, staurolite, sillimanite, and garnet, formed. Metamorphic zones can be recognized but in detail the definition of mineral isograds is complicated by the narrow width of the zones, polyphase development of the same mineral, retrograde metamorphism and tectonic deformation. Nevertheless, isograds have been drawn which correspond in the field to the first appearance of porphyroblasts, commonly andalusite or cordierite. Only the cordierite isograd has been shown in Figure 1 because the distance between this isograd and the andalusite isograd is too narrow for portrayal.

In some localities the structural-metamorphic relationships as determined from thin section studies along the cordierite isograd are readily apparent (Fig. 2). A muscovite and chlorite schistosity developed parallel to the axial plane of a second fold. Porphyroblasts of andalusite and cordierite overgrow or push the foliae apart. The following reactions are considered probable:

chlorite + muscovite + quartz = andalusite + biotite + water (R.1)

chlorite + muscovite + quartz = andalusite + biotite + cordierite + water (R.2)

A second andalusite (Fig. 4) may overgrow the andalusite developed from reactions (1) or (2), due to a continuous rise in temperature according to the following reaction:

muscovite + cordierite → andalusite + biotite

At higher temperatures sillimanite may form in place of andalusite. Sillimanite and andalusite of reaction (3) formed concurrently or in close succession (Figs. 2, 4, 5). This would place the reaction, according to Holdaway (1971) and Anderson et al. (1977) at a minimum of 500°C and at pressures below 4 kb*. No kyanite has been reported from the area and there is no reason to suppose that pressure conditions in lower amphibolite facies rocks were not consistently below 4 kb. The temperature, on the other hand, was quite variable, as illustrated by the isograds in the vicinity of Indin Lake (Fig. 1).

Granulite facies rocks have been reported from the Ghost Lake area (Folinsbee, 1940; Robertson and Folinsbee, 1974). Garnet and cordierite gneiss and schist, presumably derived from the Yellowknife Supergroup sediments, locally contain biotite-garnet-cordierite-sillimanite-hypersthene-spinel assemblages. Some of the granodiorite rocks yielding Pb-Pb ages of 2500 Ma contain inclusions of gneiss with similar granulite assemblages.

Upper amphibolite facies rocks west of Indin Lake (Fig. 1) consist principally of migmatite. A general increase in the abundance of the leucosome component of the migmatite toward the west is presumably caused by increased temperatures but the 'heat sink' effect of partial melting apparently constrained prograde metamorphism to the upper amphibolite facies in most localities. In addition to the water liberated during prograde metamorphic reactions, some water must have been introduced. Thus porphyroblasts are commonly altered by hydration to biotite and related minerals.

The Kenoran Orogeny culminated with the intrusion of essentially undeformed plutonic rocks along the southern part of the Bear-Slave boundary. The plutons were emplaced contemporaneously with uplifts of migmatitic complexes derived in part from supracrustal rocks of the Yellowknife Supergroup and from older granitic crust. The migmatites surrounding and making up parts of these granitoid terranes were either part of the late Kenoran intrusive episode or were formed during the regional metamorphism that formed the "spotted" and "nodular" schists of the region.

Detailed examination of the plutons in the Mattberry Lake area (Smith, 1966) revealed that the intrusive rocks are multi-phase and comprise a variety of compositions. The central parts of the plutons are undeformed but the margins are foliated (Fig. 6) and this foliation locally overprints marginal migmatitic layering in the host rocks. It is concluded that the host migmatites were formed during the regional metamorphism. The granitic diapirs and metamorphic complexes rose as partly solid diapirs after the peak of regional metamorphism as shown by fabric relationships, but the granitoid rocks contained residual heat which caused contact retrograde metamorphism of the host rocks. Garnet and some pegmatitic segregations were formed at this time.

Proterozoic Metamorphism

The Slave Province rocks have not undergone extensive Hudsonian deformation and thus contrast structurally with their exhumed equivalents in the Bear Province. Archean rocks of the Slave Province must have been metamorphosed and deformed during the Hudsonian Orogeny but the nature of the metamorphism, and to a lesser extent, the deformation, was such that it left little visible imprint on the rocks.

Early K-Ar determinations on biotite and muscovite from granulitic rocks and pegmatites yielded ages ranging from about 2500 Ma to 1850 Ma with histogram peaks at 2200 Ma and 1900 Ma. Results of Rb-Sr studies on minerals (McGlynn, 1972) and whole rock systems (Frith et al., 1977) concur. Proterozoic (about 2000 Ma) intrusions, although small, are widespread (Frith et al., 1977); alkaline intrusions dating about 2200 Ma are present along the Bear-Slave boundary fault (Martineau and Lambert, 1974).

Deformation in this region is more evident than metamorphism. Northwest striking cleavage is present in many rock types, particularly the metasediments. Gentle warps of strata and minor development of chlorite are locally apparent. Most deformation, however, is brittle; large scale northwest-trending sinistral oblique-slip faults have been mapped in the Indin Lake basin (Fig. 1). Diabase dykes parallel the faults. Leech's (1966) 'preferred' age of 2000 Ma for the dykes is probably too high, the age is probably closer to 1900 Ma. In any case, the faulting and dyke intrusion are approximately contemporaneous and formed as a result of regional dilation in which heavy volcanic rocks were downwarped relative to the lighter granitic areas. Brittle deformation, particularly at the boundary regions, resulted in mylonitized zones that extend for many miles. Tourmaline- and muscovite-bearing pegmatites, pegmatitic granite, and a group of high level granodiorite stocks were also intruded about this time.

* 1 kb = 1 x 10^5 kPa.

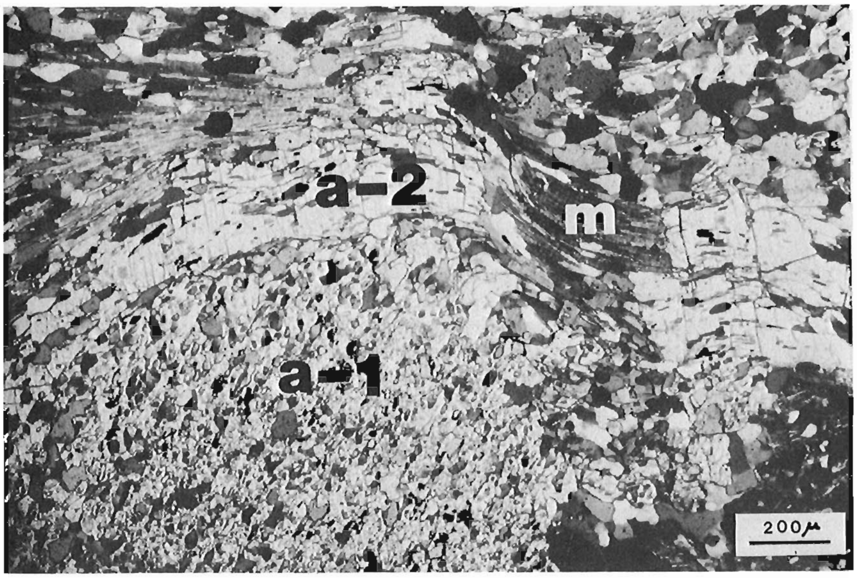

Figure 4

Poikiloblastic andalusite in Yellowknife metagreywacke (a-1) is overgrown by optically continuous second phase andalusite (a-2) which grew from the breakdown of muscovite (m) and cordierite. The poikiloblasts and groundmass is mostly quartz and plagioclase. (x-nicol)

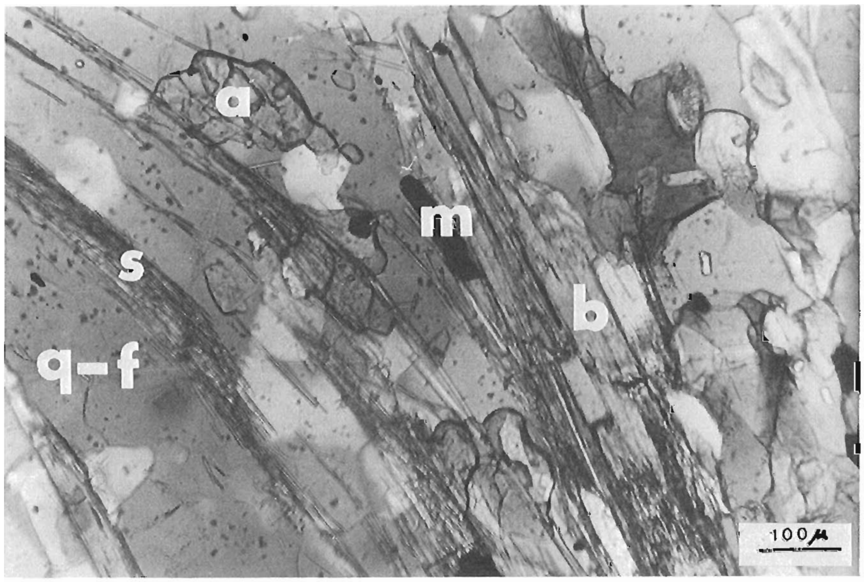

Figure 5

Fibrolite (s) has grown along with biotite (b) as reaction products of muscovite and cordierite. Andalusite (a), magnetite (m) are also present, all in a quartzofeldspathic matrix (q-f). (double exposure in plain light and x-nicols)

Metamorphic isograds assumed their current position during the 1900 Ma period of dilation and vertical adjustment, modified only slightly by recent slow-moving cratonic tilts and uplifts.

METAMORPHISM ALONG THE BEAR-SLAVE BOUNDARY

The Bear-Slave boundary was drawn along the eastern margin of the Aphebian-covered terrane (Stockwell, 1961). This boundary coincides over much of its length with faults (Fig. 6) and the eastern margin of deformational structures and textures formed after the deposition of Proterozoic cover rocks. The fault trace is obscure near Mesa Lake, on the Slave Province margin, but is pronounced toward Mattberry Lake. The fault extends into the Slave craton toward Basler Lake where it is the locus of an early and middle Proterozoic alkaline intrusive complex (Bigspruce Lake).

The boundary is barely distinguishable on the metamorphic map of the region (Fig. 1). Bear and Slave terranes in the boundary zone are each characterized by an Abukuma-type of metamorphism, principally of upper to lower amphibolite facies grade, except for a thin wedge of Proterozoic greenschist and subgreenschist rocks adjacent to the Slave craton. The Abukuma-type metamorphism in the Slave Province predated the large scale phase of granitic intrusion at approximately 2500 Ma, but postdated the deposition of supracrustal rocks (∼2700 Ma). In the Bear Province the metamorphism took place between 2200 Ma and 1800 Ma (Frith et al., 1977).

Aeromagnetic profiles across the Bear-Slave boundary reflect basement-cover relationships, particularly those within the Bear Province. Where Snare Group cover is thick, anomalously low signatures are evident (Fig. 6a). However, where the sedimentary cover is thinner the aeromagnetic pattern is influenced mostly by the basement. Curiously, the profiles near the boundary are smooth and atypical of both the Slave and Bear provinces (Fig. 6b). The homogeneous nature of the aeromagnetics also transgresses the boundaries of formations of strongly contrasting lithology. The cause of the homogenization was most likely a high grade metamorphism that predated the deposition of the Snare Group but postdated the Archean granite, metasediments and metavolcanics. A ∼2200 Ma event reset Rb-Sr whole rock systems (Frith et al., 1977) on both sides of the boundary where high grade rocks are present in folded anticlinorial or dome-like structures.

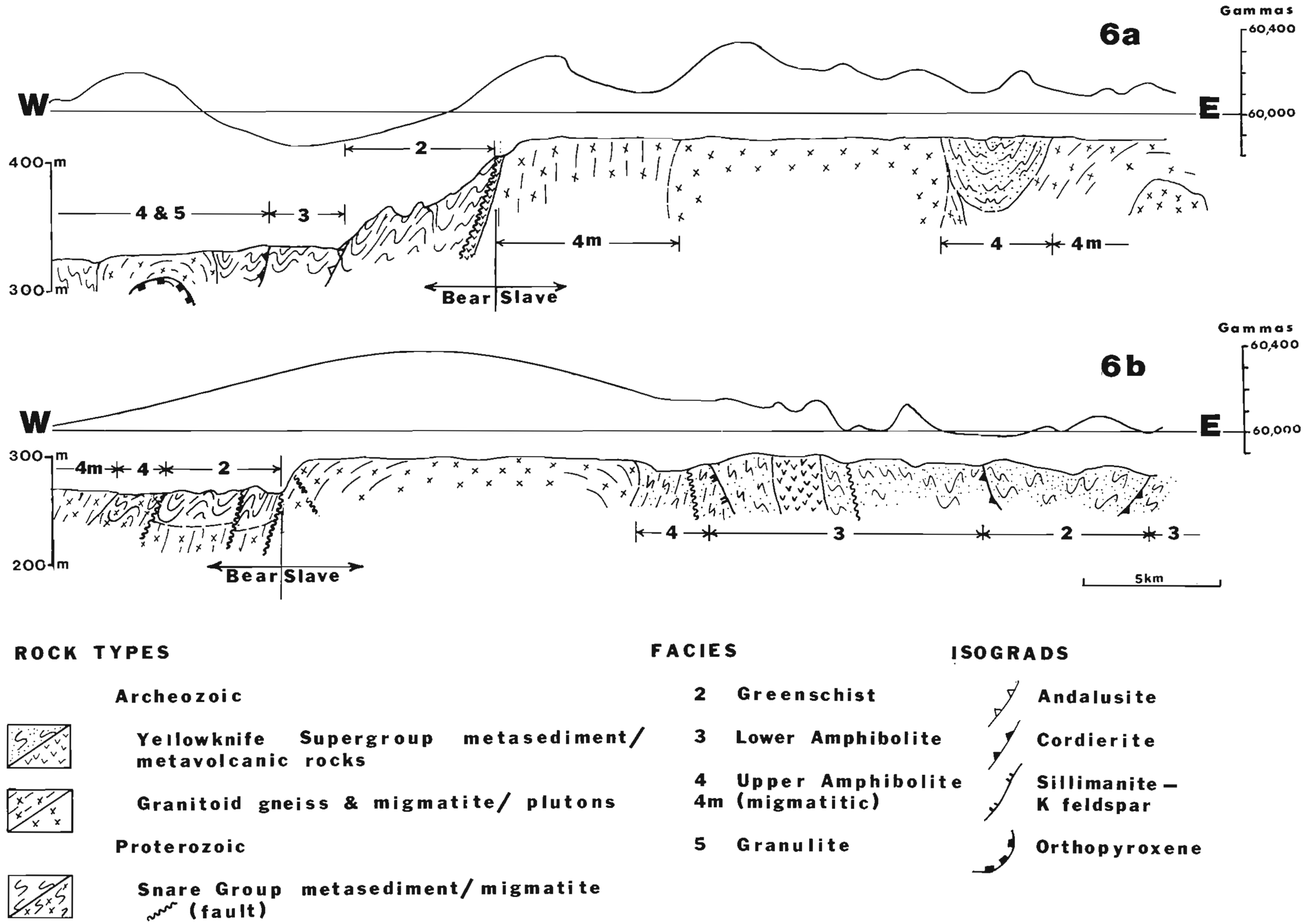

Figure 6a. The synoptic profile across the Bear-Slave boundary at Arseno Lake shows upper amphibolite rocks of the Slave craton faulted against Snare Group rocks, which prograde from greenschist to upper amphibolite facies. The Arseno Lake upright fold zone has an abnormally low aeromagnetic signature, due to a thick folded pile of low magnetic metasediments. In and around the Emile River gneiss domes, granulite facies rocks have been diapirically brought close to the surface, but the orthopyroxene isograd is related to an earlier period of metamorphism that is overprinted by 2000-1800 Ma thermal event(s).

Figure 6b. The Mattberry Lake aeromagnetic profile contrasts with Fig. 6a by being smooth and shield-like across the Bear-Slave boundary. The migmatites on the extreme west side were formed ∽2200 Ma, which compares with a similar aged diapir on the Slave side. It is considered probable that a pervasive metamorphism along the border region homogenized the magnetics at this time. The Snare Group rocks are thought to be thin and do not adversely influence the aeromagnetic profile.

METAMORPHISM AND TECTONISM IN THE BEAR PROVINCE

The Coronation Geosyncline has been divided into four tectonic belts (Hoffman, 1973) which include a platform sequence, a foreland thrust belt, a metamorphic belt, and a volcanic belt. These belts trend northerly, terminating at the Slave boundary which trends south-southeast. The southwestern Slave Province is in contact with the metamorphic belt, which was the most tectonically active part of the geosyncline (Fig. 1). The Proterozoic deformational and metamorphic events that took place along this belt had a profound effect on the Slave craton.

The metamorphic belt in contact with the southwestern Slave Province may be divided into three tectonic zones, each with its characteristic structural and metamorphic signature (Fig. 6).

The Arseno Lake Upright Fold Zone

The fold geometry of Snare Group rocks has been discussed in detail by Ross and McGlynn (1965). The rocks extend for about 85 km from Mesa Lake to Basler Lake. They are mostly isoclinal folds, upright to westerly-dipping, that plunge at shallow angles toward the north end of the Arseno Lake sedimentary basin. The metamorphism increases in grade towards the west from greenschist near the Bear-Slave boundary, to amphibolite grade and granulite grade toward the west. The single phase of deformation becomes polyphase more or less at the cordierite-sillimanite isograd which defines the western limit of this fold zone. Near the Slave craton greenschist facies shale becomes schistose within 5 km of the craton. Porphyroblasts of biotite and muscovite are the first metamorphic minerals to appear, then garnet and staurolite, then andalusite and cordierite.

The Emile River Gneiss Domes

The region to the west of the andalusite-cordierite isograd (Fig. 1) has undergone polyphase deformation and remobilization of the Archean basement into several gneiss domes. The structural geology of this region has been studied in detail (Frith and Leatherbarrow, 1975). The structural style of folding is best illustrated by the conglomeratic marker unit that blanketed the southwest part of the Arseno Lake map area. In essence the stratified rocks have been deformed by three phases of folding. The first (F_1) is the same as that observed in the Arseno Lake zone. Around the gneiss domes F_1 is locally refolded by F_2. The gneiss domes formed during F_2 folding. A concentric gneissosity in the gneiss dome is probably contemporaneous.

The metamorphism of this region is generally of upper amphibolite grade, but in the Emile River-Arseno Lake area granulite facies grade assemblages are present. Like the style of deformation, the metamorphism is sequential. The first phase is the highest, as shown by the presence of spinel and retrograded orthopyroxene. The age of the granulite is unknown, but in the Ghost Lake area relationships suggest that the granulite facies is Archean (see below). Stable assemblages are cordierite-sillimanite-garnet and biotite. Mg-Fe distribution coefficients between cordierite and garnet suggest a load pressure of 3.5 kb at a temperature of 630°C (Nielson, 1978). The garnet occurs as porphyroblasts which are commonly elongated or spiralled by the second fabric (S_2). Other garnets are idioblastic and contain inclusion trails that parallel S_2. Similarly, sillimanite needles parallel bedding or S_1, but a second coarse sillimanite overprints the S_1 fabric. Also, leucosome segregations in the gneiss are folded by S_2 but porphyroblasts of orthoclase overgrow the S_1 fabric. All of these relationships suggest a two-stage development of metamorphic minerals.

Interpretation of the structural-metamorphic data can assume two distinct metamorphic episodes, or a longer, continuous one through changing P-T conditions. The isotopic data for the region suggest a two-stage evolution (Frith et al., 1977): an early event at $\backsim$2200 Ma, and a later, longer-lived event between 2000 and 1800 Ma.

Porphyroblastic batholiths

Granodioritic bodies and the surrounding gneiss and schist make up the third zone of the metamorphic belt. The nature of the contacts between the granodiorite bodies and their host rocks is confusing. In places the contacts are concordant, in others, discordant. Gradational contacts are also known and in isolated occurrences unconformable relationships have been recognized. The understanding of these contact relationships is essential to understanding the tectonics and metamorphic history of the basement rocks. A synopsis of the various types of contacts is presented below.

Unconformable relationships

Lord (1942) studied an unconformable basement-cover relationship along the present Bear-Slave boundary in the Mattberry Lake area. He found essentially undeformed Snare Group rocks unconformably overlying granitoid basement. This locality and many others were re-examined along the Bear-Slave boundary and within the Bear Province itself (Frith, 1973). At one locality Snare Group quartz pebble conglomerate was unconformably deposited in an erosional scour formed in basement granodiorite (Frith et al., 1977). Both the conglomerate and the porphyroblasts in the basement were deformed by the deformational event that characterizes the Snare Group rocks of the area (Ross and McGlynn, 1966). At the unconformity, typical porphyroblastic granodiorite is locally transformed into augen gneiss (Frith et al., 1977).

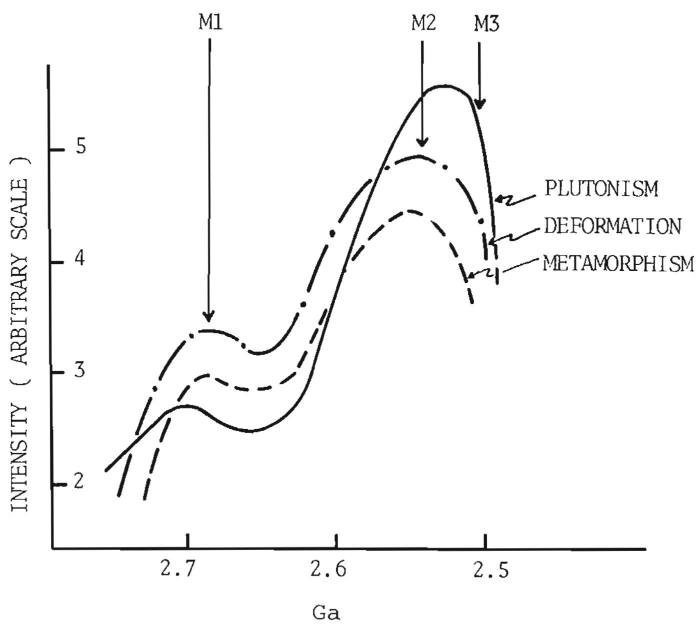

Figure 7. The relationship of metamorphism, deformation and plutonism of the "Kenoran" Orogeny in southwestern Slave Province is shown with an arbitrary intensity scale and time in Ga years. M_1 generally reaches greenschist, M_2 reaches upper amphibolite, but M_3 is a contact metamorphic episode which is regionally retrogressive. M_2 is closely related to deformation. Generally the peak of metamorphism follows the peak of deformation, but locally, the reverse takes place.

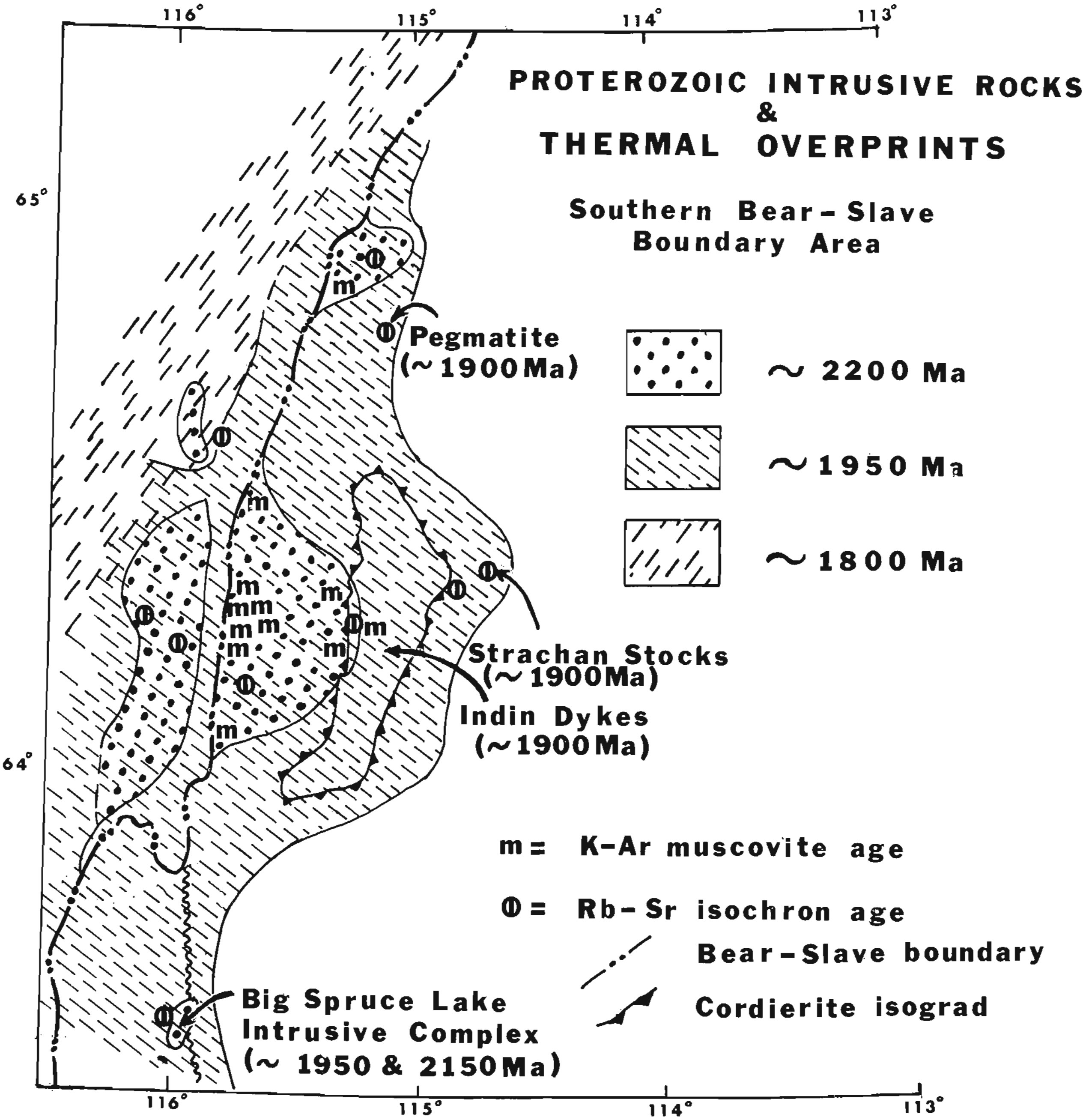

Figure 8. An interpretation of the Proterozoic metamorphic history of the southern Bear-Slave border region. The metamorphism is thought to be related to the rift break-up of the Slave craton into circum-Slave sedimentary basins. Metamorphism and minor intrusion took place at 2200 and 1950 Ma in the Slave Province. The last ~1800 Ma event is part of the "Hudsonian" Orogeny, but it represents the close of orogenic activity that only affected the western part of this area.

Cross-cutting relationships

The east margin of the batholith west of Emile River is vertical to inward-dipping. Along the boundary, dyke-like apophyses of alaskitic vein material, containing porphyroblasts of K feldspar, locally intrude the gneiss derived from the Snare Group rocks. However, these same dykes also intrude the batholith, suggesting that they were derived from the body of the batholith itself.

Faulted contacts

The steep margins of the batholiths may be faults, the batholith having moved upward relative to the host sediments. Streaked out remnant conglomerate cobbles of the base of the Snare Group locally plunge steeply along the faults. This suggests that the fault contact is a displaced, steeply dipping unconformity. The faulting is assumed to be associated with late (1800-1700 Ma) upward movement of basement rock.

Gradational contacts

Some contacts with the batholith and their host rocks are gradational. Microcline porphyroblasts within the host sediments at the margin of the batholith and within the batholith are similar. Some porphyroblasts in the batholith exhibit rapakivi texture, thought to have formed during a metasomatic interval that postdated the initial porphyroblastic growth. It is this stage of metasomatism that produced the feldspar growth within the host sediments. These porphyroblasts are undeformed, suggesting that the thermal metasomatic event was the last major tectonic event to take place.

SYNTHESIS

The southwestern Slave Province and marginal Bear Province region was affected by four thermal events:

1. A pre-"Kenoran" event at 3000 Ma.
2. The "Kenoran" Orogeny waned by 2500 Ma.
3. An early Proterozoic event at 2200 Ma.
4. The "Hudsonian" Orogeny at 2000-1800 Ma.

1. Pre-Kenoran metamorphism was eclipsed by the more pervasive metamorphism that occurred during the Kenoran Orogeny. The pre-Kenoran granitic basement was preserved from isotopic homogenization only where the Kenoran-aged metamorphic grade did not exceed greenschist (or possibly lower amphibolite grade). The low grade rocks were mylonitized and altered by pegmatitic introduction so that little evidence bearing on their early metamorphic history remains. The $\backsim$160 Ma difference between Rb-Sr whole-rock and zircon concordia ages, and the presence of metasomatic textures, suggest that the $\backsim$3000 Ma Rb-Sr date represents a metamorphic event involving the introduction of K and Rb.

2. The Kenoran Orogeny occurred in two phases. The early phase was dominated by deformation and magmatism related to dilational instability that produced large scale volcanism and related intrusive activity along the margins of deep basins in a granitic crust. Sedimentation was contemporaneous with basin formation. The sediments were folded along axes that roughly parallel the basin margins, and the volcanic rocks that occur along the margins were stretched in a vertical direction. The rocks were metamorphosed during subsidence (M_1) but there is no evidence to suggest that this metamorphism exceeded the greenschist facies on a regional scale.

The main phase of orogeny that followed, overshadowed all previous deformation and metamorphism. This phase was of long duration and the tectonic style was one of compression rather than dilation. Magmatism was largely plutonic rather than volcanic, and a high grade of regional metamorphism (M_2) and a tighter style of isoclinal folding prevailed. The overall sequence of events was: deformation, metamorphism and plutonism. However, there was considerable overlap of these in time (Fig. 7), so that in some places porphyroblasts may be overprinted by a deformation fabric. Plutonic uplift was nevertheless consistently later than both deformation and metamorphism and the latent heat of the diapirs caused marginal, commonly retrograde, metamorphism (M_3) of the host rocks. A regional metamorphism prevailed during this phase of the orogeny and as a result most Slave craton K-Ar ages do not exceed 2500 Ma.

The highest grade of metamorphism attained was the granulite facies. A belt of metasediments in the Ghost Lake area contains biotite-garnet-cordierite-sillimanite-hypersthene-spinel assemblages. Similar assemblages are present in inclusions found in a 2500 Ma granitoid pluton (Robertson and Folinsbee, 1974) which establishes an Archean (M_2?) age for the granulite facies metamorphism.

3. The 2200 Ma metamorphic event that affected the southern Bear-Slave boundary region is outlined in Figure 8. The evidence for this event is mainly isotopic (Frith et al., 1977). Metamorphic ages (K-Ar, muscovite and Rb-Sr, pegmatites) have been obtained from gneiss domes, anticlinorial structures, and retrograde Archean volcanic rocks (Fig. 8). In addition, some parts of the Big Spruce syenite-carbonatite complex were intruded at this time along a fault trace that parallels the Bear-Slave boundary (Fig. 8). This early metamorphism and fault-related magmatism, is most likely connected to early rift formation of the Coronation Geosyncline. This period of dilational tectonics was followed by postdepositional (Snare Group) compressional tectonics that culminated in orogeny. The uplift of high grade metamorphic terranes took place considerably later, possibly as late as 1800 Ma. Rocks on each side of the Bear-Slave boundary were raised to different levels. On the Bear Province side, northerly trending anticlinorial and doming structures were formed. The amount of uplift reflects the depth of burial necessary for upper amphibolite and granulite facies metamorphism. On the Slave Province side, where granulite facies is less common, uplift was not generally as great.

4. The $\backsim$1950 Ma event was characterized by structural downwarping, faulting and intrusion of dykes in the Indin Lake basin, accompanied by the intrusion of the Strachan stocks along the Indin Lake basin-granitoid basement contact (Frith et al., 1977). The dykes are presumed to be contemporaneous with sills and dykes emplaced within the Bear Province. A cleavage parallel to the northwest dykes, occurs in Slave craton rocks. Chlorite may be present along the cleavage surfaces.

Granitic rocks along the Bear-Slave contact yield K-Ar biotite ages that cluster on histogram plots at 1900 Ma. Metamorphism represented by these ages is presumed to be the same as the greenschist metamorphism of the Snare Group rocks (Lord, 1942). The metamorphism increases westward from greenschist through the sillimanite isograd (Fig. 1). In the vicinity of Emile River, however, gneiss domes, polyphase deformation, migmatization and the monotonous composition of the subgreywacke, preclude recognition of metamorphic zonation.

Metamorphism in the westernmost part of the area is associated with batholith emplacement and dates about 1800 Ma (Fig. 8), overprinting the 1900 Ma thermal event that is more prevalent near and along the margins of the Slave craton. The 1800 Ma event was the last metamorphism to affect the region west of the sillimanite isograd. It is enigmatic whether or not the 1800 and 1900 Ma events are one and the same. If both the 1900 Ma and 1800 Ma events are considered as a single event, the 1900 Ma ages would reflect cessation of metamorphism near the Slave craton, but continued metamorphism toward the centre of the metamorphic belt. Figure 8 depicts the eastern limits on the Slave craton. The metamorphism along the central metamorphic zone, perhaps because of higher heat flow, would have been longer lived. A second thermal pulse which involved batholithic development would have overlapped earlier metamorphism.

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METAMORPHISM OF THE ARSENO LAKE AREA, NORTHWEST TERRITORIES

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Nielsen, P.A., Metamorphism of the Arseno Lake area, Northwest Territories; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 115-122, 1978.

Abstract

Metamorphic grades in a suite of aluminous Proterozoic Snare Group metasediments and polymetamorphic remobilized pre-Hudsonian (Archean?) paragneisses from the Arseno Lake area range from low in the chlorite zone adjacent to the Bear-Slave Province boundary in the eastern part of the area, to high in the cordierite-almandine-K feldspar zone in the west. Six isograds have been mapped: biotite, andalusite, cordierite, sillimanite, sillimanite + K feldspar, and almandine + K feldspar ± cordierite. This sequence corresponds to the 'low-pressure' facies series of Miyashiro.

Microprobe analyses of the ferromagnesian silicates show that the scale of equilibirum in the cordierite-almandine-K feldspar zone is of the order of 1 mm. This small scale equilibrium is most probably a function of the CO_2-rich nature of the metamorphic fluid.

The P-T values of metamorphism in the Arseno Lake area range from ∽350°C @ 2.0 kb (2 x 10^5 kPa) where X_{H_2O} = 1.0 in the chlorite zone to ∽630°C @ 3.2 kb (3.2 x 10^5 kPa) with $X_{H_2O} \leq 0.5$ for the cordierite-almandine-K feldspar zone.

Résumé

Les degrés de métamorphisme observés dans une suite de métasédiments alumineux du groupe de Snare, d'âge protérozoïque, et des paragneiss pré-hudsoniens polymétamorphiques remobilisés (d'âge archéen?), rencontrés dans la région du lac Arseno, passent graduellement d'une zone chloritique proche de la limite entre la province de l'Ours et la province des Esclaves dans la partie est de la région, à une zone à cordiérite, almandine et feldspath potassique à l'ouest. En s'appuyant sur la distribution spatiale des assemblages minéraux observés en lame mince, on a cartographié les isogrades suivantes.

1. *Biotite*
2. *Andalousite*
3. *Cordiérite (disparition de la muscovite et de la chlorite)*
4. *Sillimanite (disparition de l'andalousite)*
5. *Sillimanite-feldspath potassique (disparition de la muscovite et du quartz)*
6. *Almandine-feldspath potassique ± cordiérite (disparition de la biotite, de la sillimanite, et du quartz).*

Cette succession d'assemblages minéraux correspond à la série de faciès de "basse pression" de Miyashiro (1961).

L'analyse par microsonde des silicates ferromagnésiens indique que l'échelle d'équilibre de la zone à cordiérite-almandine et feldspath potassique est d'environ 1 mm. Cette échelle d'équilibre réduite est très probablement dépendante de la richesse en CO_2 du fluide métamorphique.

Les valeurs de P et T du métamorphisme qui caractérisent la région du lac Arseno varient entre environ ∽350°C @ 2.0 kb là où X_{H_2O} = 1.0 dans la zone chloritique et ∽630°C @ 3.2 kb avec X $X_{H_2O} \leq 0.5$ pour la zone à cordiérite-almandine et feldspath potassique.

INTRODUCTION

The Arseno Lake area is in the Bear Province, 260 km north-northwest of Yellowknife, Northwest Territories, and extends from 115°30'W to 116°00'W and 64°15'N to 64°45'N (Fig. 1). East of the Bear-Slave Province boundary, bedrock consists of the Yellowknife Group and associated intrusive rocks of Archean age (Lord, 1942; McGlynn and Ross, 1963; Ross and McGlynn, 1965). West of the boundary, bedrock comprises predominantly metamorphosed Proterozoic Snare Group sediments, with widely distributed occurrences of remobilized Archean ortho- and paragneisses (Lord, 1942; Frith et al., 1974; Frith and Leatherbarrow, 1975; Leatherbarrow and Frith, 1975; Nielsen, 1975, 1976, 1977). K-Ar radiometric dating by the Geological Survey of Canada (Wanless et al., 1965, 1966, 1968, 1970) established an age of 1815 Ma for the metamorphism of the Snare Group. Rb-Sr dating reported by Frith et al. (1974) established a minimum age of 2712 ± 89 Ma for the granitoid core of the gneiss dome in the western part of the area.

GENERAL GEOLOGY

Most of the area is underlain by low to intermediate grade metamorphosed sedimentary rocks of the Snare Group (Lord, 1942). Siliceous dolomite at the base of the sequence is overlain by interbedded thin bands of siltstone and shale, subgreywacke, rusty pyritic shale that in places contains quartzite horizons, quartz pebble conglomerate, and calcareous argillite.

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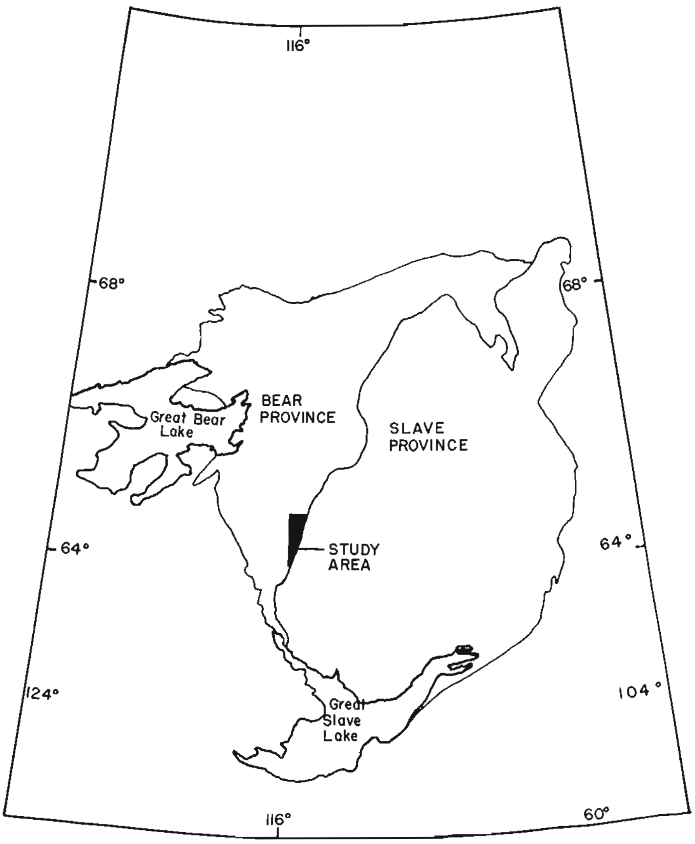

Figure 1. Index map of the Arseno Lake area.

Table 1

Comparison of the central Abukuma Plateau and the Arseno Lake area

Central Abukuma Plateau		Arseno Lake Area
chlorite + muscovite	increasing metamorphic grade ↓	chlorite + muscovite
biotite + muscovite		biotite + muscovite
spessartine-rich garnet		spessartine-rich garnet
andalusite + muscovite		andalusite + muscovite
cordierite + muscovite		cordierite + muscovite
sillimanite + muscovite		sillimanite + muscovite
sillimanite + K feldspar		sillimanite + K feldspar
almandine-rich garnet		almandine-rich garnet
		almandine garnet + K feldspar ± cordierite
10-20 km wide		15 km wide
1.0-4.0 kb	P range	2-3.5 kb
300-800°C	T range	350-650°C
Igneous Association		
abundant granite abundant in situ migmatite		minor granite scarce in situ migmatite

Higher grade metamorphic rocks of the area include local pegmatite, minor injection migmatite, and remobilized Archean granitoid rocks that occur with retrograded pre-Hudsonian paragneisses. The latter are similar to the Archean paragneisses in the Ghost Lake area described by Folinsbee (1940, 1941, 1942).

Basement to the Snare Group metasediments are the Yellowknife Group metasediments and metavolcanics, and locally Kenoran intrusive rocks. Throughout much of the central area, only metamorphosed Snare Group sediments are exposed. Basement in the western part of the area consists of remobilized Archean granitoid rocks and pre-Hudsonian paragneisses, which together define the core of the southwesterly plunging gneiss dome shown in Figure 2.

Deformation of the Snare Group differed from east to west, and the cordierite isograd approximates the boundary between two distinct structural domains. Summaries of structures in the eastern domain were compiled by McGlynn and Ross (1953), Frith and Leatherbarrow (1975), and Frith (1978). Structural data for the western domain were compiled by Frith and Leatherbarrow (1975) and Frith (1978).

METAMORPHISM

The metamorphism of the Proterozoic Snare Group and remobilized pre-Hudsonian paragneiss was a 'low-pressure' type event similar to that in the Abukuma Plateau (Miyashiro, 1958). Table 1 presents a comparison of the Abukuma Plateau and the Arseno Lake area. Metamorphic grade increases from the chlorite zone of the greenschist facies at the base of the Snare Group sediments in the east to the cordierite-almandine-K feldspar zone of the granulite facies in the west. First appearance of minerals or mineral assemblages observed in thin section define the following isograds:

BIOTITE ISOGRAD	–	Biotite
ANDALUSITE ISOGRAD	–	Andalusite
CORDIERITE ISOGRAD	–	Cordierite (muscovite + chlorite out)
1ST SILLIMANITE ISOGRAD	–	Sillimanite (andalusite out)
2ND SILLIMANITE ISOGRAD	–	Sillimanite + K feldspar (muscovite + quartz out)
ALMANDINE + K FELDSPAR ISOGRAD	–	Almandine + K feldspar ± cordierite (biotite + sillimanite + quartz out)

Locally, the first appearance of garnet occurs between the biotite and andalusite isograds. These low grade garnets occur only in Mn-rich bands. Commonly observed mineral assemblages are summarized in Table 2. The areal distribution of mineral assemblages and isograds is illustrated in Figure 3a and a metamorphic facies map in Figure 3b.

Petrology

Chlorite zone

The lowest grade metamorphic rocks occur immediately west of the Bear-Slave Province boundary and extend from 0.2 km to as much as 1.2 km into the Bear Province. The samples from this zone contain chlorite, white mica, quartz, and detrital feldspars, as well as ilmenite and minor authigenic pyrite. These are very fine grained rocks comprising quartz and plagioclase clasts set in a matrix of recrystallized quartzofeldspathic material and interleaved chlorite and white mica.

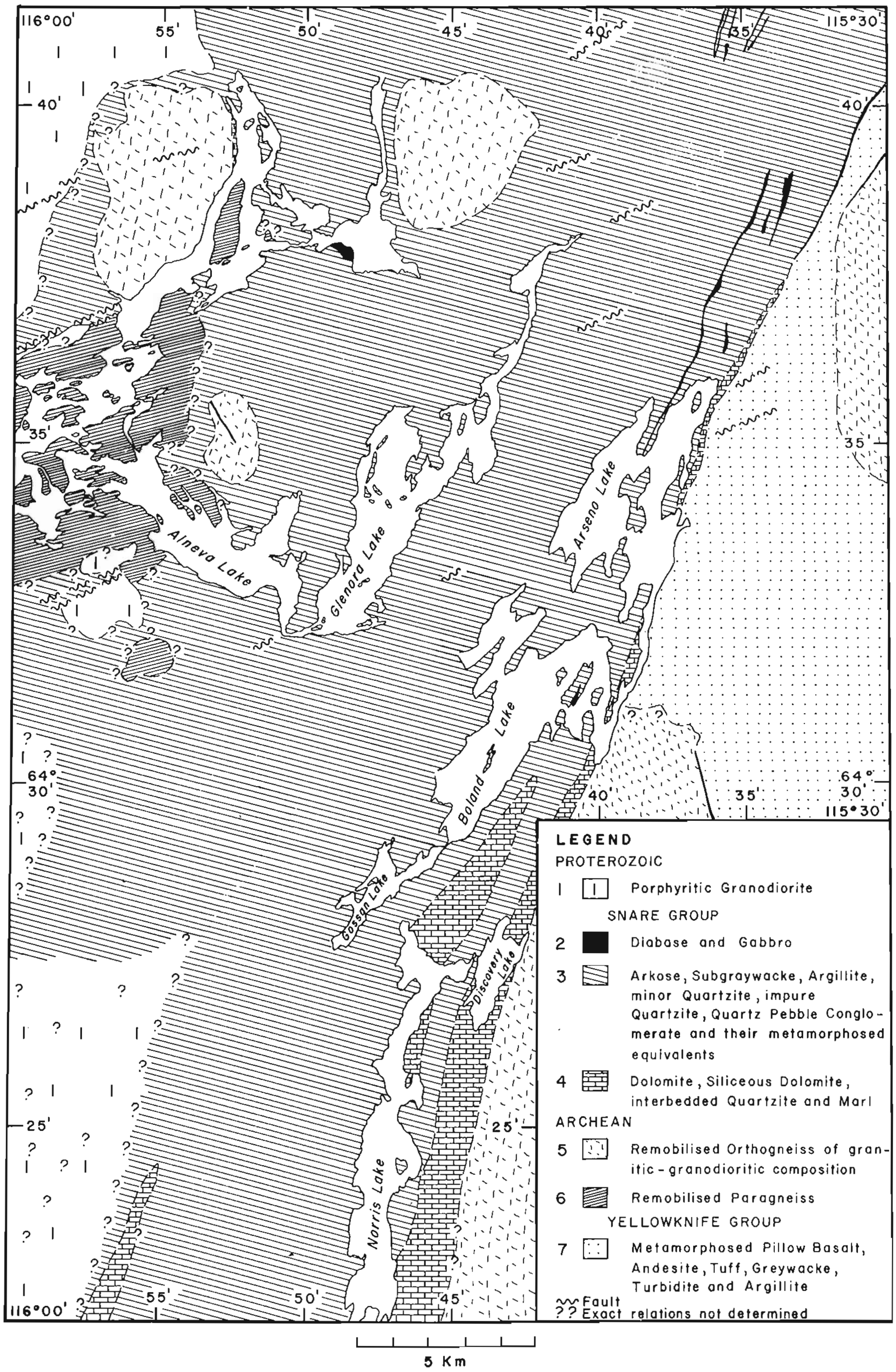

Figure 2. Generalized geological map of the Arseno Lake area. Data compiled from Lord (1942), Frith and Leatherbarrow (1975) and Nielsen (1977).

Biotite zone

The biotite zone ranges in width from 5.5 km east of the gneiss dome in the northern part of the area, to 8 km in the area south of the gneiss dome. The biotite zone is defined by the first appearance of biotite resulting from reaction (1):

$$\text{chlorite} + \text{muscovite}_{(1)} + \text{ilmenite} \rightarrow \text{biotite} + \text{muscovite}_{(2)} + \text{quartz} + \text{rutile} + H_2O \pm \text{K feldspar} \qquad (R.1)$$

This reaction was proposed by Ramsay (1973a,b,c) for rocks of similar composition and setting. Muscovite$_1$ is phengitic, whereas muscovite$_2$ is closer to the ideal muscovite formula. The major role of muscovite, however, is to provide K for biotite. K feldspar may or may not appear, depending on whether K and Al are present in excess of that required by biotite and muscovite$_2$.

Samples from the lowest part of the biotite zone are characterized by a phyllitic or lepidoblastic arrangement of fine grained chlorite and white mica in a matrix of granoblastic quartz and feldspar. Large ragged biotite porphyroblasts that cross this foliation commonly contain a network of rutile needles. Ilmenite, graphite and minor pyrite are also present. In hand specimen, the rocks are best described as spotted biotite phyllites.

In the upper part of the biotite zone, biotite is commonly interleaved with chlorite or muscovite, and where abundant it defines the foliation.

In the uppermost part of the biotite zone, andalusite appears as large porphyroblasts (up to 3 cm long) and most of the quartz and plagioclase have recrystallized to an equilibrium texture characterized by triple point junctions and straight line boundaries. Spessartine-rich poikiloblastic garnets are present in rocks of the appropriate composition.

Table 2

Commonly observed mineral assemblages from individual thin sections

GREENSCHIST FACIES

Chlorite Zone

- chlorite-sericite-quartz-plagioclase-ilmenite ± detrital K feldspar and pyrite
- calcite-dolomite-quartz-tremolite

LOWER AMPHIBOLITE FACIES

Biotite Isograd

Biotite Zone

- biotite-sericite-chlorite-quartz-plagioclase-ilmenite-rutile ± detrital K feldspar and pyrite
- biotite-muscovite-spessartine garnet-quartz-plagioclase-rutile-ilmenite
- biotite-muscovite-quartz-plagioclase-ilmenite-rutile

Andalusite Isograd

Andalusite Zone

- biotite-muscovite-andalusite-quartz-plagioclase-ilmenite-rutile
- biotite-chlorite-andalusite-quartz-plagioclase-ilmenite-rutile

Cordierite Isograd

Cordierite Zone

- biotite-muscovite-cordierite-andalusite-quartz-plagioclase-rutile-ilmenite
- biotite-cordierite-andalusite-quartz-plagioclase-ilmenite-rutile
- biotite-cordierite-fibrolite-andalusite-plagioclase-ilmenite-rutile

1st Sillimanite Isograd

- biotite-cordierite-sillimanite-fibrolite-plagioclase-ilmenite-rutile
- biotite-cordierite-muscovite-orthoclase-plagioclase-ilmenite-rutile-quartz

UPPER AMPHIBOLITE FACIES

2nd Sillimanite Isograd

- biotite-cordierite-sillimanite-microcline-quartz-plagioclase ± ilmenite ± rutile

GRANULITE FACIES

Almandine-K feldspar Isograd

- biotite-cordierite-sillimanite-microcline-quartz-plagioclase-garnet ± ilmenite ± rutile
- biotite-cordierite-garnet-microcline-quartz-plagioclase ± ilmenite ± rutile
- biotite-garnet-microcline-quartz-plagioclase ± ilmenite ± rutile

Retrograded Archean Paragneiss

- biotite-cordierite (spinel-sillimanite)[1] K feldspar-plagioclase-quartz ± ilmenite ± rutile
- biotite-cordierite (spinel)[1] K feldspar-plagioclase-quartz ± ilmenite ± rutile
- biotite-cordierite (spinel)[1] garnet-K feldspar-plagioclase-quartz ± ilmenite ± rutile
- biotite-cordierite (spinel-sillimanite)[1] garnet-K feldspar-plagioclase-quartz ± ilmenite ± rutile
- biotite-cordierite-garnet-orthopyroxene-K feldspar-plagioclase-quartz ± ilmenite ± rutile
- biotite-orthopyroxene-plagioclase-quartz-K feldspar ± ilmenite ± rutile
- calcite-serpentine (forsterite)[1] ilmenite

[1] Minerals within brackets are found only as inclusion within the preceding mineral in the table.

Table 3

Pressure-temperature conditions of metamorphism for the Arseno Lake area

Mineral zone	Temperature[1] range (°C)	Pressure[2] range (kb)	X_{H_2O}[3]	Controls and Restrictions[4]
Chlorite	≤ 350	≃ 2	1.0	chl+mus/bio
Biotite	350 – 530	2.0 – 2.5	1.0	as triple point; py/as+qu on as; chl+mus+qu/crd+bio on as
Cordierite Andalusite	530 – 560	2.5 – 3.0	≥0.9	as triple point; chl+mus+qu/crd+bio on as; as boundary
Cordierite Sillimanite	560 – 600	3.0 – 3.2	≥0.8	chl+mus+qu/crd+bio on as; mus+qu/S+ksp
Sillimanite K feldspar	600 – 650	3.1 – 3.3	≥0.7	mus+qu/S+ksp; bio+S+qu/crd+ksp
Almandine Cordierite K feldspar	≥ 630	≥ 3.2	<0.5	bio+S+qu/crd+ksp; bio+S+qu/crd+gar+ksp; crd/gar+S+qu

[1] Temperatures are ± 50°C.

[2] Pressures are ± 500 bars

[3] Where X_{H_2O} <1, the probable error is ± 0.2

[4] Abbreviations are the same as in Figure 4. The sequence of minerals is REACTANTS/PRODUCTS

Cordierite zone

The cordierite zone (Fig. 3a) ranges in width from 0.9 km in the south to 9.4 km in the north. It can be divided into 3 subzones according to the aluminous mineral assemblages present:

the cordierite-andalusite assemblage,

the cordierite-sillimanite-muscovite assemblage, and

the cordierite-sillimanite-K feldspar assemblage

The beginning of the cordierite-andalusite assemblage is defined by the first appearance of cordierite as a product of reaction (2):

chlorite + muscovite + biotite$_1$ + quartz + ilmenite → cordierite + biotite$_2$ + andalusite + rutile + quartz (R.2)

Biotite$_2$ is richer in Ti and Mg and depleted in total Al and Fe relative to biotite$_1$.

The transition from the cordierite-andalusite assemblage to the cordierite-sillimanite assemblage is based on the inversion of andalusite to sillimanite. The cordierite-forming reaction would no longer have chlorite as a reactant, and sillimanite would replace andalusite as a product.

The 2nd sillimanite isograd defines the transition from the cordierite-sillimanite-muscovite assemblage to the cordierite-sillimanite-K feldspar assemblage. The cordierite-forming reaction (3) at the 2nd sillimanite isograd is:

biotite$_1$ + muscovite + quartz + Mg (in solution) → biotite$_2$ + cordierite + sillimanite + K feldspar + rutile + H_2O (R.3)

Biotite$_2$ is lower in total Al and Mg and shows an enrichment in Ti and Fe. This trend towards Al-depletion and Fe-enrichment in biotite continues above the 2nd sillimanite isograd by reaction (4):

biotite$_1$ + sillimanite + quartz ± ilmenite → cordierite + K feldspar + biotite$_2$ + H_2O ± rutile (R.4)

Cordierite-andalusite assemblage rocks are nodular, containing large diffuse cordierite poikiloblasts and euhedral andalusite porphyroblasts. Chlorite and muscovite are mutually exclusive whereas biotite is ubiquitous and abundant. The rocks containing this assemblage are more massive than those in the biotite zone, and cleavage and a well-defined foliation are less common except where biotite is a major component. Cordierite-sillimanite-muscovite assemblage rocks are similar in appearance to cordierite-andalusite assemblage rocks except that the former are rusty weathering. Cordierite-sillimanite-K feldspar assemblage rocks are massive to gneissose, nodular, and are also rusty weathering. The nodules comprise porphyroblasts of cordierite, bundles of fine grained sillimanite (variety fibrolite), and large prisms of coarsely crystalline sillimanite.

Cordierite-almandine-K feldspar zone

The cordierite-almandine-K feldspar zone, which is approximately 9 km in diameter (Fig. 3a) represents the highest grade of metamorphism recorded in the Snare Group metasediments. The first appearance of garnet + K feldspar ± cordierite arising from reactions (5) or (6) defines the base of the zone:

biotite$_1$ + sillimanite + quartz + ilmenite → garnet + bioite$_2$ + K feldspar + rutile + H_2O (R.5)

Biotite$_2$ is richer in Mg and Ti than biotite$_1$.

biotite$_1$ + sillimanite + quartz + ilmenite → garnet + cordierite + biotite$_2$ + K feldspar + rutile + H_2O (R.6)

Here, as in reaction (5), biotite$_2$ is richer in Mg and Ti than biotite$_1$.

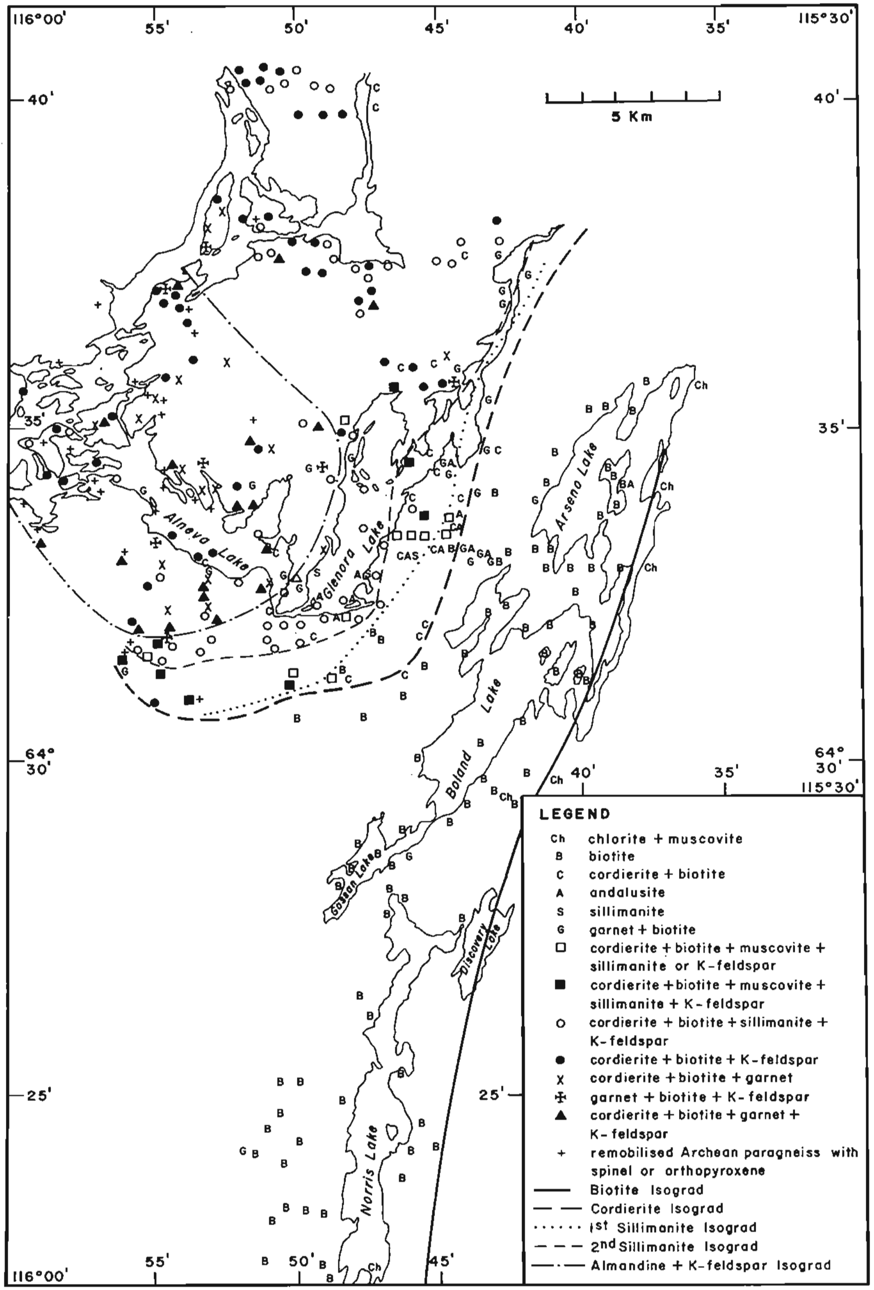

Figure 3a: Map showing the distribution of key mineral assemblages and isograds.

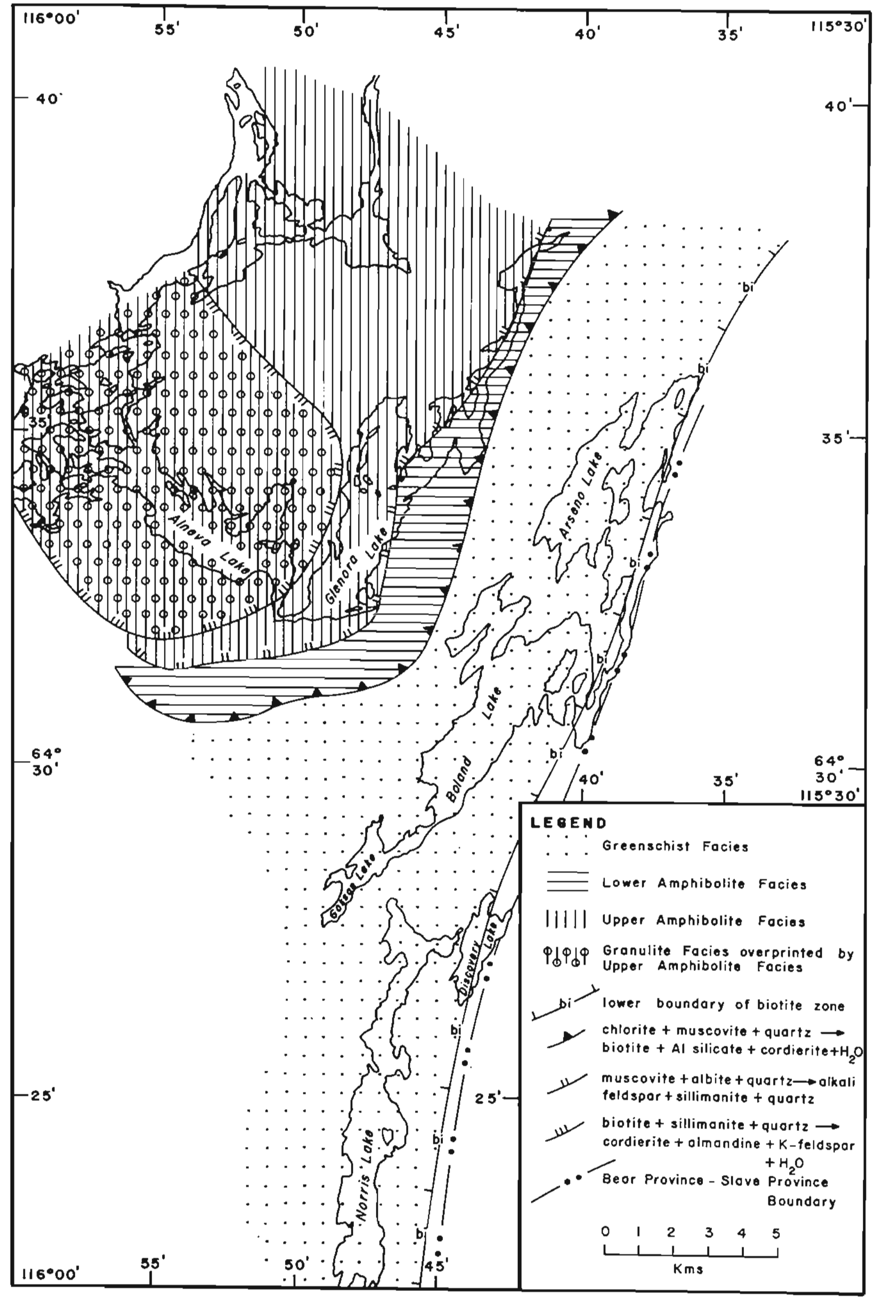

Figure 3b: Metamorphic facies map of the Arseno Lake area. The granulite facies is defined by the coexistence of almandine with K feldspar; the remaining facies are defined according to Winkler (1968).

Reaction (5) apparently proceeded in rocks with a low Al/Fe+Mg ratio, and resulted in cordierite-free assemblages. In rocks with a higher Al/Fe+Mg ratio, reaction (6) formed the common assemblage biotite + cordierite + garnet + K feldspar.

The stability field of cordierite was not exceeded and the assemblage garnet + sillimanite + quartz was not observed. The absence of this reaction places an upper limit on the load pressure (of metamorphism) in the Arseno Lake area.

REMOBILIZED PRE-HUDSONIAN PARAGNEISS

Polymetamorphic rocks exposed in the western part of the Arseno Lake area were retrograded to the cordierite-almandine-K feldspar zone during the Hudsonian metamorphism, but relict grains of hercynite-spinel solid solution and orthopyroxene persist. The occurrence of spinel enclosed by cordierite, the occurrence of cordierite + coarsely crystalline sillimanite in rocks of the cordierite-sillimanite-muscovite zone through to the cordierite-almandine-K feldspar zone, in addition to the occurrence of embayed orthopyroxene, lead to the conclusion that these gneisses were originally metamorphosed under much higher P-T conditions.

CONDITIONS OF METAMORPHISM

The facies series described above, characterized by low pressure and a high thermal gradient, suggests that the pressure of metamorphism ranged from about 2 kb[1] in the biotite zone to 3.5 kb in the cordierite-almandine-K feldspar zone, and the temperature ranged from ∽350°C in the biotite zone to ∽650°C in the cordierite-almandine-K feldspar zone. Figure 4 shows the experimentally determined equilibria used to define the P-T trajectory for the Arseno Lake area. Table 3 summarizes the P-T-X_{H_2O} conditions for each mineral assemblage zone.

ACKNOWLEDGMENTS

The data presented here are the results of field and laboratory work done at the Department of Geology, University of Alberta, as part of a doctoral dissertation. I wish to acknowledge the direct financial support of the Boreal Institute for Northern Studies of the University of Alberta and the Department of Indian and Northern Affairs, Yellowknife. Indirect support from the National Research Council of Canada to staff members in the Department of Geology, most notably in support of the microprobe laboratory by grants to D.G.W. Smith, and for field work by grants to R.S.J. Lambert, are gratefully acknowledged. The research benefited from many discussions with R.S.J. Lambert, D.G.W. Smith, and M.J. Holdaway.

[1] 1kb = 10^5 kPa.

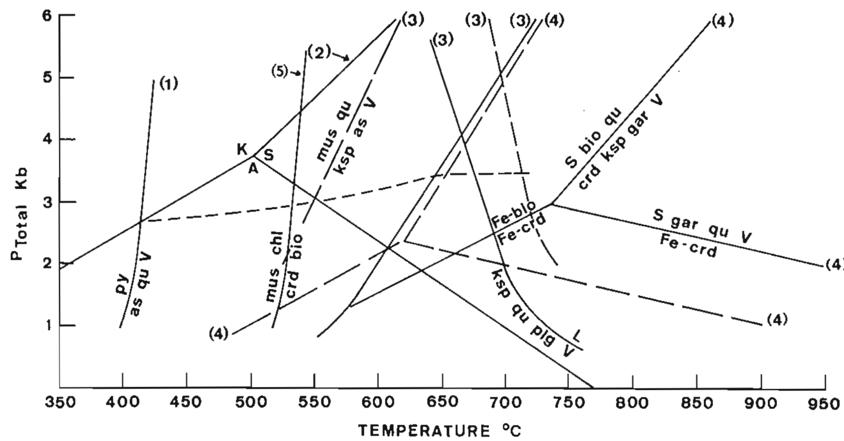

Figure 4. Diagram showing the experimentally determined and calculated equilibria used to define the P_{total} – T conditions of metamorphism in the Arseno Lake area. The dotted line shows the inferred P-T trajectory.

(1) Haas and Holdaway (1973); (2) Holdaway (1971);
(3) Kerrick (1972); (4) Holdaway and Lee (1977);
(5) Winkler (1968)

py – pyrophyllite, as – aluminum silicate,
qu – quartz, K – kyanite,
A – andalusite, S – sillimanite,
mus – muscovite, chl – chlorite,
crd – cordierite, bio – biotite,
ksp – K feldspar, gar – almandine,
plg – plagioclase, v – water vapour,
L – melt.

Dashed lines are for conditions of X_{H_2O} <1.0.

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METAMORPHISM OF THE ATHABASCA MOBILE BELT
A SUBSURFACE EXTENSION OF THE CHURCHILL PROVINCE

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Burwash, R.A., Metamorphism of the Athabasca Mobile Belt, a Subsurface Extension of the Churchill Province; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 123-127, 1978.

Abstract

Metamorphic processes in the subsurface Precambrian of western Canada are documented in the Athabasca Mobile Belt (which extends southwest from Lake Athabasca) using multivariate analysis of polymetamorphic textures as recognized in thin section, combined with modal analyses, major element geochemistry and trace heavy metal content.

Two sequential but overlapping types of Hudsonian metamorphism acted on the Kenoran crystalline complex and its Aphebian cover. Widespread brittle deformation of the crystalline complex is documented in three stages: (1) formation of epidote and chlorite, with destruction of biotite and iron sulphides and loss of Ni and Zn; (2) formation of secondary foliation, with destruction of plagioclase and loss of Hg; and (3) formation of quartz and hematite, with destruction of plagioclase, K-feldspar and iron sulphides.

Prograde regional metamorphism accompanied by K-Rb metasomatism, concurrent with or later than the cataclastic processes, resulted in the pervasive growth of microcline blasts, myrmekitic replacement of plagioclase by quartz and microcline, and growth of new biotite at the expense of chlorite, leucoxene and epidote. Growth of biotite and closure of the K-Ar and Rb-Sr isotopic systems, as indicated by geochronologic studies in northeastern Alberta, ended at about 1800 Ma.

Résumé

Les processus métamorphiques dans le Précambrien qui se trouve sous le Phanérozoïque dans l'Ouest du Canada sont reconnus dans la zone mobile de l'Athabasca et qui s'étend vers le sud-ouest à partir du lac Athabasca jusqu'aux Rocheuses. L'identification de textures polymétamorphiques en lames minces, combinée avec l'analyse modale, la géochimie des éléments majeurs et la concentration des métaux lourds en trace fournissent les données pour une analyse multivariée.

Deux types de métamorphisme hudsonien consécutifs, mais se recoupant, ont affecté le complexe cristallin kénoranien et sa couverture aphébienne. Une déformation cassante du complexe cristallin est largement répandue et établie suivant ces trois étapes:

(1) formation d'épidote et de chlorite, avec destruction de la biotite et des sulfures de fer et, perte de Ni et Zn,

(2) formation d'une foliation secondaire, avec destruction du plagioclase et perte de Hg,

(3) formation de quartz et d'hématite, avec destruction du plagioclase, du feldspath potassique et des sulfures de fer.

Le métamorphisme régional prograde, accompagné par un métasomatisme du K et du Rb, est contemporain ou postérieur aux processus cataclastiques. Ce métamorphisme et ce métasomatisme ont causé la croissance et l'infiltration de blastes de microcline, le remplacement myrmékitique du plagioclase par du quartz et de la microcline et la croissance de nouvelle biotite aux dépens de la chlorite, de la leucoxène et de l'épidote. La croissance de la biotite et la fermeture des systèmes isotopiques K-Ar et Rb-Sr se sont achevées il y a environ 1 800 Ma, comme l'indiquent des études géochronologiques détaillées dans le nord-est de l'Alberta.

INTRODUCTION

Petrographic studies by J. Krupička in 1968-69 of core samples from the subsurface Precambrian of western Canada resulted in the recognition of numerous examples of reworked crystalline basement rocks. Textural studies, combined with modal analyses (Burwash et al., 1964), previously determined isotopic ages, and major element geochemistry led to the concept of the "Athabasca Mobile Zone" (Burwash and Krupička, 1969, 1970). Finally application of multivariate analysis and comparisons of trend surface maps with inferred tectonic boundaries of major structural units of the western Canadian Shield by Burwash and Culbert (1976) led to a redefinition in a more restricted sense of the "Athabasca Mobile Belt".

From a number of possible methods of multivariate analysis, Culbert found four to be useful: R-mode analysis, with trend surface mapping; stepwise discriminant analysis; rotated discriminant analysis; and canonical trend surface mapping. Burwash and Culbert (1976) gave a condensed interpretation of the petrologic significance of the factors derived from the various types of analyses. In this paper the textural criteria of Krupička and the principal factors derived from multivariate analyses are examined with more specific reference to probable metamorphic processes and the sequence in which these processes may have acted.

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CLASSIFICATION OF REWORKED ROCKS

The reworking of crystalline rocks involves two processes: brittle deformation and recrystallization. They may occur sequentially or simultaneously. In Krupička's scheme of classification (Burwash and Krupička, 1969), qualitative scales are established for both processes. Four levels of deformation are recognized:

- 0 – no recognizable deformation
- 1 – strained rocks
- 2 – cataclastic rocks
- 3 – mylonitic rocks

The assessment of recrystallization is more subjective than that for deformation, and three levels are assigned:

- 0 – without recrystallization
- 1 – with partial recrystallization
- 2 – with complete recrystallization

With the combination of these two scales, 12 classes of reworked rocks are possible (Fig. 1). In fact, two of the classes, (0,1) and (0,2), were not recognized in our sample suite. Krupička (pers. comm.) also recognized four other classes (Fig. 1):

[R] – reworked rocks with recrystallization outlasting brittle deformation, the final level of deformation now indeterminate.

$[R]^1, [R]^2, [R]^3$ – reworked rocks with late stage brittle deformation of level 1, 2, or 3 following recrystallization.

The path of the processes which can be recognized and described qualitatively by thin section observation is to the right and downward on Figure 1. The arrows represent various possible sequences of reworking. The solid double line may approach the "normal" path of reworking. With increasing recrystallization the degree of brittle deformation becomes more difficult to judge. Class [R] is thus placed below classes for which recrystallization is level 2. Rocks of classes $[R]^1$, $[R]^2$, and $[R]^3$, associated with late shearing, represent a reversal of the trends of recrystallization.

APPLICATION OF STEPWISE DISCRIMINANT ANALYSIS

The progressive changes in mineralogy, bulk chemistry, and texture taking place during reworking are clearly shown by stepwise discriminant analysis (Burwash and Culbert, 1976). Before discussing these changes, a documentation of the sample population and variables is warranted.

Starting with 183 unweathered basement core samples, we eliminated undeformed diabases and acid volcanic rocks, quartzites, and amphibolites. The diabases and acid volcanics were inferred to be post-Hudsonian and therefore not related to the problem under study. The quartzites and amphibolites are definitely part of the Hudsonian metamorphic complex. However, their mechanical behaviour during brittle deformation is distinctly different from that of the average gneiss of granodioritic composition. Their inclusion in the discriminant analysis might tend to obscure the trends of the "average gneissic basement". With the deletions mentioned above, the sample population was reduced to 143. Their distribution among the various classes of reworked rocks is shown in Figure 2. Class [R] rocks were reassigned to class (2,2). Rocks of classes $[R]^1$, $[R]^2$ and $[R]^3$ were reassigned to classes (1,2), (2,2) and (3,2) respectively (Fig. 1).

A total of 31 variables which were entered into file for stepwise discriminant analysis, are: weight per cent SiO_2, TiO_2, Al_2O_3, Fe_2O_3, CaO, Na_2O, K_2O; specific gravity; volume per cent K-feldspar, plagioclase, quartz, biotite, chlorite, epidote, iron oxides, iron sulphides, leucoxene, sphene; parts per million Rb, Sr, Cu, Mo, Pb, Zn, Co, Ni, Ag, Hg; Rb/K; and presence or absence of myrmekite and foliation. The only deletion which might have proved useful was modal percentage of white mica (muscovite plus sericite).

For the purposes of analysis the classes of reworked rocks shown on Figure 2 were divided into five groups, with a limited overlap of two groups. The groups are:

Group 1	(n=55),	class (0,0), "undeformed".
Group 2	(n=17),	classes (1,0) and (1,1), "slightly deformed".
Group 3	(n=23),	classes (2,0) and (2,1), "strongly deformed".
Group 4	(n=15),	classes (3,0), (3,1) and (3,2), "intensely deformed".
Group 5	(n=45),	classes (1,2), (2,2), (3,1) and (3,2), "recrystallized".

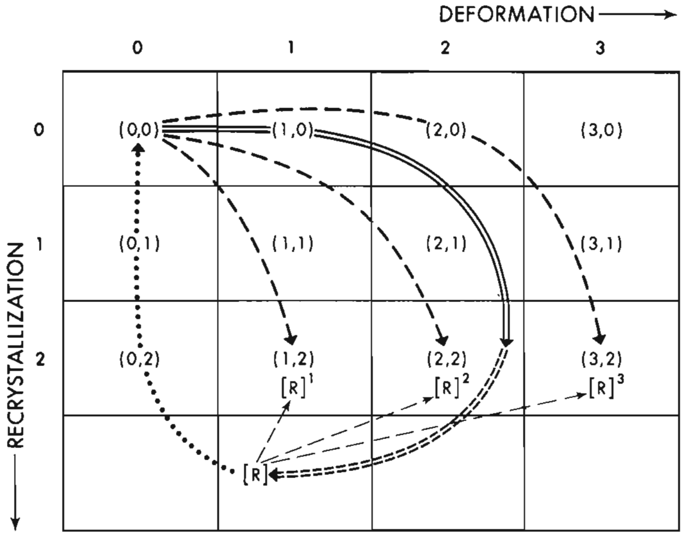

Figure 1. Classes of reworked rocks as defined by qualitative scales of deformation and recrystallization. Single dashed lines represent possible paths of deformation and recrystallization. The heavier double solid and dashed lines represent varying degrees of certainty for the "average" path of quartzo-feldspathic gneisses. The dotted line, which completes the cycle, is inferred but was not observed in thin section.

RECRYSTALLIZATION ↓ / DEFORMATION →	0	1	2	3
0	(0,0) 55	(1,0) 10	(2,0) 6	(3,0) 3
1	(0,1) —	(1,1) 7	(2,1) 17	(3,1) 3
2	(0,2) —	(1,2) 4	(2,2) 29	(3,2) 9

Figure 2. Number of samples of reworked rocks in each class from the subsurface Precambrian of western Canada (n = 143).

The stepwise discriminant analysis then proceeds in four steps between successive pairs of groups. At the final step all samples not included in Group 5 are combined under the heading of "unrecrystallized" rocks. The results of the analysis are given in Figure 3.

PROGRESSIVE STEPS IN REWORKING OF CRYSTALLINE ROCKS

In the classic concepts of "progressive regional metamorphism" an underlying assumption is the rapid attainment of equilibrium with change in conditions of temperature and pressure. In the reworking of crystalline rocks the criteria used to define the classes depend largely on the persistence of unstable relicts, either mineralogical or textural to indicate the past history of the rock. A quartzofeldspathic rock, with unimodal grain-size distribution, which has reached equilibrium by recrystallization is unlikely to be recognizable as a "reworked" rock.

A composite picture of the changes occurring during each step of reworking is provided by combining the descriptive classification of Krupička (Burwash and Krupička, 1969, p. 1382-83) and the ratios of intergroup variance to total variance given in Figure 3.

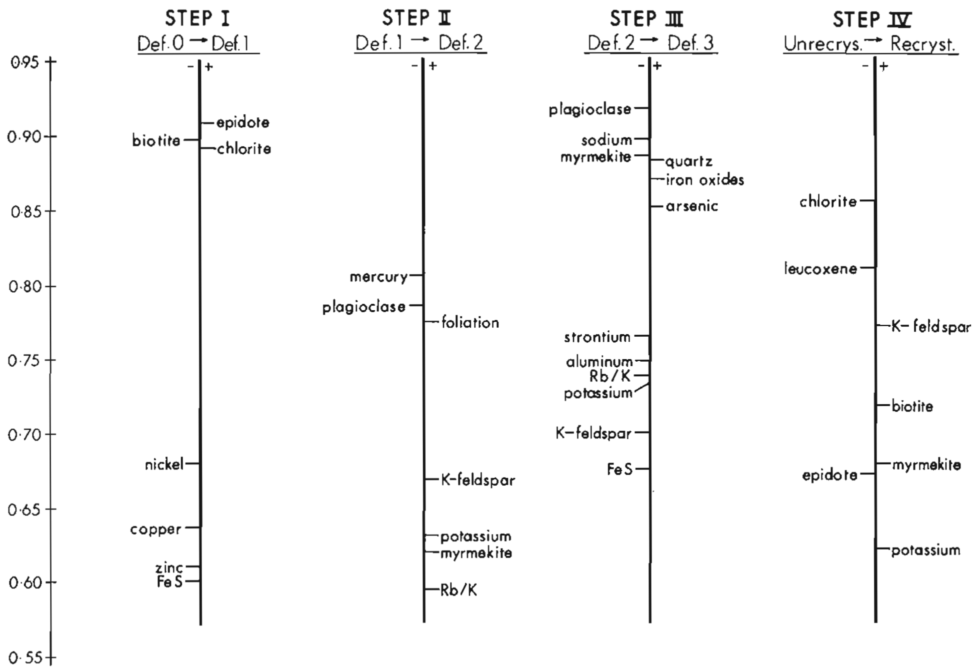

Figure 3. Steps in the progressive reworking of crystalline rocks from the subsurface Precambrian of western Canada. Scale at left is the ratio of inter-group variance to total variance as determined by stepwise discriminant analysis. Positive and negative signs represent increase and decrease in variable.

Step I ("Undeformed" to "slightly deformed")

(1) The predeformational textures are still easily discernable although plagioclase twin lamellae are warped and quartz is partly crushed.

(2) Sparsely distributed, narrow zones of strained or sheared minerals are developed.

(3) In these narrow zones, biotite is destroyed with concurrent formation of epidote and chlorite. The chlorite interlayers the warped biotite or completely replaces it as pseudomorphs. The epidote commonly recrystallizes as finely granular aggregates along the narrow shear zones.

(4) The heavy trace metals Ni, Cu and Zn are lost, along with the destruction of iron sulphides.

Step II ("Slightly deformed" to "strongly deformed")

(1) The predeformational texture is largely obliterated with all minerals xenomorphic.

(2) Potassium feldspars survive and even grow with the initial stages of recrystallization and metasomatism which accompany most deformation of this intensity. The porphyroclasts are surrounded by a mortar of granulated quartz and strongly deformed micas and hornblende.

(3) Hiatal grain-size distribution is typical, with bands and lenses of crushed minerals imparting a megascopic foliation to the rock.

(4) Plagioclase is mechanically crushed and altered to epidote and white mica.

(5) The decrease in mercury is indicative of the continued loss of heavy trace metals.

(6) Most reworked rocks pass directly from Step II to Step IV without going through Step III. For this reason, the "typical path of deformation" (Fig. 1) is drawn through classes (2,0) and (2,1).

Step III ("Strongly deformed" to "intensely deformed")

(1) Continued crushing leads to a uniformly fine grained cataclasite, with only scattered relict potassium feldspar porphyroblasts.

(2) Destruction of plagioclase and potassium feldspar, coupled with increase in quartz causes an apparent decrease in sodium, aluminum and potassium.

(3) Myrmekite, which is normally developed along the margins of plagioclase grains, is destroyed.

(4) In hand specimens these rocks are often reddish brown in colour. The increase in iron oxide reflects continuing destruction of iron sulphide.

(5) Our data base unfortunately did not distinguish ferrous from ferric iron. Thin section observations suggest that most of the iron oxide is now in the form of hematite.

(6) The increase in arsenic implies the action of hydrothermal solutions. The widespread association of quartz veins with highly sheared rocks in many areas of the exposed Canadian Shield suggests that there has been a metasomatic addition of silica and arsenic in Step III.

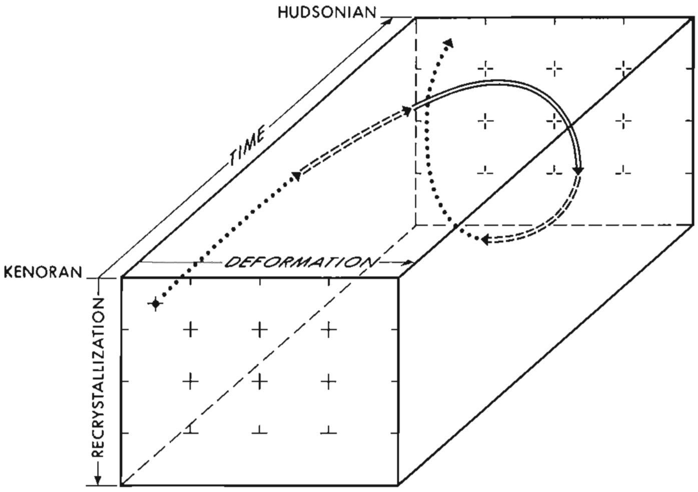

Figure 4. Inferred path of deformation and recrystallization for the Athabasca Mobile Belt as a whole. The legend for the "average" path is the same as Figure 1. The interval of time is from the close of the Kenoran Orogeny to the close of the Hudsonian Orogeny (700 Ma).

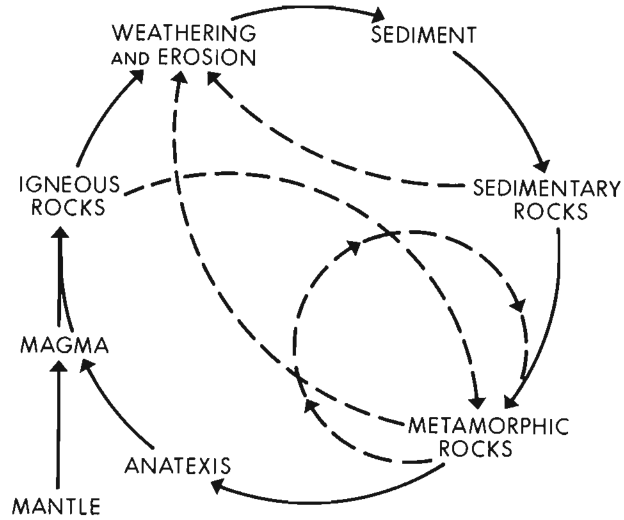

Figure 5. The rock cycle, showing the restricted path (small circle) of the reworked rocks in the Athabasca Mobile Belt. Kenoran igneous rocks (left centre) are reworked with Kenoran metamorphic rocks (lower right) to become parts of the gneissic complex by the close of the Hudsonian.

Step IV ("Unrecrystallized" to "recrystallized")

(1) The textural criteria for recognizing recrystallization are the healing of old crystals, the growth of new crystals of old minerals and the growth of new minerals.

(2) In this study the development of new microcline porphyroblasts is the dominant process of recrystallization. The unusual abundance of microcline and myrmekite and the great variability of their forms, coupled with the dynamometamorphic texture, characterize the polymetamorphic rocks of the mobile zone (Burwash and Krupička, 1969).

(3) All schemes of multivariate analysis reported by Burwash and Culbert (1976) showed potassium feldspar and/or potassium as the most important variables in factors related to recrystallization.

(4) Chlorite and epidote which formed in Step I are now progressively destroyed with the growth of biotite. A petrographic characteristic of many of the gneisses of the Athabasca Mobile Belt is the presence of remarkably fresh, well crystallized biotite.

(5) Leucoxene, which had formed during Steps I, II and III with the destruction of sphene and titaniferous magnetite now disappears. Most of the titanium probably goes into rutile inclusions in the new, deep brown biotite.

TEMPORAL ASPECTS OF REWORKING

The application of the K-Ar method of age determination to mica separates from the western Churchill Province fixes the time of crystallization (or recrystallization) of biotite and its closure to subsequent argon loss at approximately 1800 Ma (Burwash et al., 1962; Baadsgaard and Godfrey, 1972). Studies using the Rb-Sr and U-Pb methods on whole rock samples and mineral separates from northwestern Saskatchewan (Koster and Baadsgaard, 1970) and northeastern Alberta (Baadsgaard and Godfrey, 1972) indicate relict tectonic enclaves of Kenoran igneous and metamorphic rocks within the Hudsonian complex. The record of at least two major orogenic events (Kenoran and Hudsonian) in the Lake Athabasca area is relevant to the interpretation of the history of the Athabasca Mobile Belt. The regional tectonic fabric of the Athabasca Mobile Belt strongly supports its continuity with the adjacent exposed Shield (Burwash and Culbert, 1976).

The metamorphic processes observed and/or inferred from basement cores can be interpreted within the conceptual framework of Hudsonian events superimposed on Kenoran crystalline rocks. Combining the three parameters – deformation, recrystallization, and time – as rectangular co-ordinates produces Figure 4. The "average" path of deformation versus recrystallization from Figure 1 when extended through time produces a metamorphic spiral. In terms of the interval of approximately 700 Ma between the Kenoran and Hudsonian orogenies, most of the reworking probably occurred between 2200 and 1800 Ma (Burwash, 1976).

Rock-forming processes and changes in rock masses have often been represented by diagrams similar to that shown in Figure 5. The path of deformation and recrystallization plotted on Figures 1 and 4 constitutes the small circle tangential to the large circle at the words "metamorphic rocks". Igneous rocks from earlier cycles can also be incorporated into this path. In the Athabasca Mobile Belt the limited number of recognizable metasedimentary rocks found suggests that only the keels of belts of infolded Aphebian cover are preserved. Most of the gneisses studied from drill cores are believed to be Kenoran igneous or metamorphic rocks reworked during the Hudsonian. With respect to the rock cycle shown in Figure 5 we are dealing with a system with only limited input of new crustal material.

ACKNOWLEDGMENTS

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METAMORPHISM IN THE CANADIAN SHIELD OF NORTHEASTERN ALBERTA

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Abstract

A Precambrian polymetamorphic complex of igneous, metamorphic, and sedimentary rocks in northeastern Alberta reveals three distinct phases of metamorphism. During the earliest phase, Archean migmatitic para- and ortho-gneisses were metamorphosed under granulite facies conditions. In the second phase, probably related to remobilization during the Hudsonian Orogeny, rocks of the Archean migmatitic complex were subjected to amphibolite facies conditions. Some foliated granitoid rocks contain minerals characteristic of both the granulite and the amphibolite facies, but others contain minerals characteristic only of the amphibolite facies. During the third phase of metamorphism, in late Aphebian time, minor sedimentary rocks and all pre-existing rocks were metamorphosed under greenschist facies conditions. Primary structures are preserved in the low grade, minor sedimentary rocks. Parts of all major rock units show the effects of cataclasis. A regional mylonitic belt, associated with greenschist metamorphism and developed principally during the Aphebian Era, is largely restricted to the Archean gneissic complex. Unmetamorphosed Helikian sedimentary rocks of the Athabasca Formation lie unconformably on the older crystalline rocks.

Résumé

Un complexe polymétamorphique précambrien, composé de roches ignées, métamorphiques et sédimentaires, et situé dans le nord-est de l'Alberta, présente trois phases distinctes de métamorphisme. Pendant la phase la plus ancienne, des paragneiss et orthogneiss migmatitiques archéens ont été métamorphisés, dans les conditions du faciès des granulites. Au cours de la seconde phase, probablement liée à la remobilisation survenue pendant l'orogenèse de l'Hudsonien, les roches du complexe migmatitique de l'Archéen ont été soumises aux conditions du faciès des amphibolites. Certaines roches granitoïdes feuilletées contiennent des minéraux caractéristiques à la fois du faciès des granulites et du faciès des amphibolites, mais d'autres contiennent des minéraux qui ne sont caractéristiques que de faciès des amphibolites. Pendant la troisième phase de métamorphisme, vers la fin de l'Aphébien, peu de roches sédimentaires et toutes les roches préexistantes ont été métamorphisées dans les conditions du faciès des schistes verts. Les structures primaires ont été préservées dans quelques roches sédimentaires faiblement métamorphisées. Certaines sections de toutes les unités rocheuses importantes présentént une texture cataclastique. Une zone mylonitique régionale associée au métamorphisme des schistes verts, et surtout formée pendant l'Aphébien se limite en majeure partie au complexe gneissique de l'Archéen. Des roches sédimentaires non métamorphisées, de l'Hélikien et appartenant à la formation d'Athabasca, recouvrent en discordance les roches cristallines les plus anciennes.

INTRODUCTION

The Canadian Shield of northeastern Alberta is at the western edge of the exposed Churchill Structural Province (Fig. 1), and consists of massive to foliated granitoid rocks, granite gneisses, and metasedimentary rocks. In a region principally south of Lake Athabasca these rocks are unconformably overlain by Athabasca Formation sandstones, and, where the Precambrian surface gently slopes to the southwest, by Devonian sedimentary rocks (Figs. 2 and 3).

Although the dominant metamorphic episode shown by K-Ar dates on mica and hornblende records a Hudsonian thermal event at 1800 ± 50 Ma, Rb-Sr zircon age dating studies reveal a much older Archean history (2524 ± 27 Ma) (Baadsgaard and Godfrey, 1967, 1972). The geological history is characterized first by the granulite facies, later by the amphibolite facies, and finally by greenschist facies. Their distribution is shown in Figures 4 and 5.

The critical observations used in this reconstruction are based on the well exposed geology on the north side of Lake Athabasca because extensive bedrock outcrops are lacking south of the lake. The northern area has been divided into two lithostructural domains, the Leland Lake domain and the Andrew Lake domain (Fig. 2). The sequence of geological events in each is similar. However, the expression and preservation of some phases in the history differs in that the Leland Lake domain now appears as a predominantly granitoid terrane, whereas the Andrew Lake domain is largely a supracrustal granite-gneiss terrane with subordinate granitoid components.

GEOLOGICAL AND METAMORPHIC HISTORY

The geological history of the area can be subdivided into three principal tectonic episodes, which are correlated with three distinct metamorphic phases (Table 1).

Phase I

Archean crustal conditions point to the presence of a supracrustal paragneissic complex with an infrastructure of probable granitoid composition. The complex has been mapped from latitude 60°N, west of Andrew Lake, as far south as Fort Chipewyan (Figs. 1,2). It consists of granite-gneisses (biotitic and hornblendic), minor bands of high grade metasedimentary rocks, small bands and lenses of amphibolite, local agmatitic masses, and widespread migmatite (Fig. 8C). Numerous bands of metasedimentary rocks within

[1] Alberta Research Council

Table 1

Table of geologic events for the Canadian Shield, northeastern Alberta, including metamorphic history

AGE	Phase	PRINCIPAL EVENTS		LELAND LAKE DOMAIN Rocks	LELAND LAKE DOMAIN Facies	ANDREW LAKE DOMAIN Rocks	ANDREW LAKE DOMAIN Facies
HELIKIAN			Hinged Uplift			Athabasca Formation	Unmeta-morphosed
APHEBIAN	III	Deformation – retrogr. metam. – northerly fold axes Faulting – mylonitization – metamorphism	Supracrustal rocks eroded	Warren Fault mylonite	Greenschist	Allan Fault mylonite	Greenschist
		Deposition				Waugh Lake sedimentation, volcanism	
	II	Deformation – northerly fold axes Deformation – easterly fold axes Plutonic intrusion metamorphism	Uplift	Arch Lake Granite	Amphibolite	Minor granitoid plutons, Wylie Lake & Colin Lake granitic complexes	Amphibolite
		Plutonic intrusion metamorphism		Slave Granite	Amphibolite overprint	Minor granitoid plutons	Amphibolite overprint
ARCHEAN	I	Deformation – metamorphism – northerly foliation		Su – Paragneissic complex In – Granitoid rocks Remobilized	Granulite	Su – Paragneissic complex In – Granitoid rocks Remobilized	Granulite

Ongoing deformation ↑ Su – superstructure In – infrastructure Intruded ←

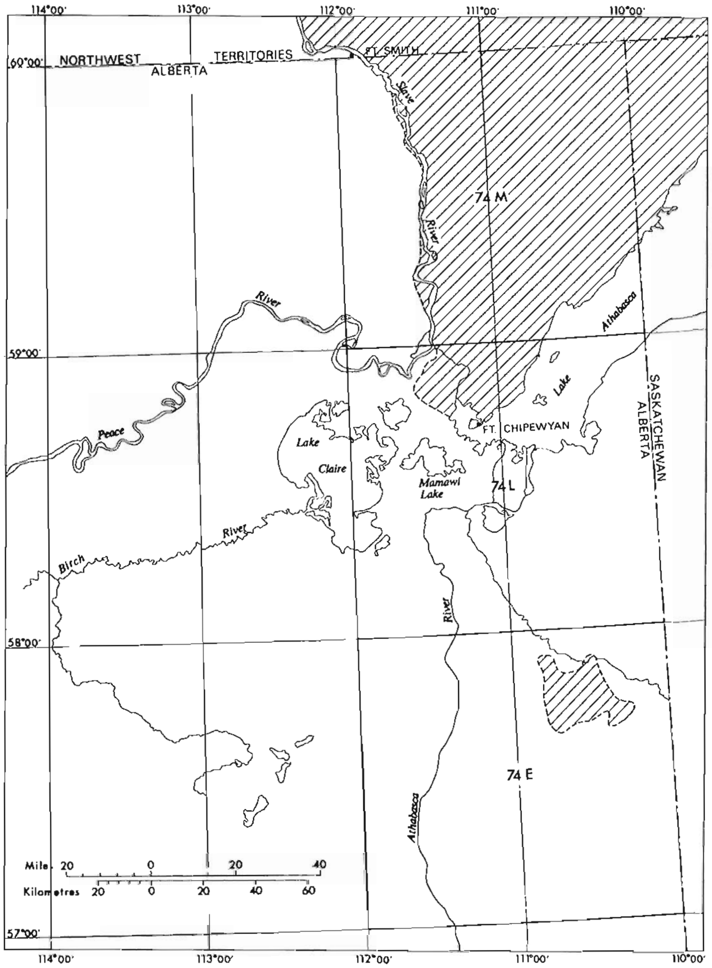


Figure 1. Location map showing the exposed Canadian Shield north and south of Lake Athabasca.

the gneisses (Fig. 8A) contain high grade metamorphic minerals, including hypersthene, sillimanite, and green spinel, indicative of granulite facies conditions (Turner, 1968).

The accompanying infrastructure probably was represented by granitoid masses such as Slave Granite[1] and minor granitoid plutons, for example, the Thesis Lake Granite. Both of these rock types are charnockitic as they contain hypersthene, and hence formed under granulite facies conditions. Sillimanite and green spinel are also typical of the major masses of Slave Granite.

South of Lake Athabasca the exposed Shield is dominated by granitoid rocks (Fig. 3), some of which reveal relicts of granulite facies conditions. High grade metasediments and granite gneisses (supracrustal rocks) are virtually absent.

Phase II

Remobilization of infrastructure materials (accompanied by reconstitution of supracrustals) during metamorphism appears to have been responsible for their emplacement higher in the crust during the Aphebian Era.

Both the Slave Granite and the Thesis Lake Granite show the effects of an amphibolite facies overprint (Figs. 6A, B, D and F) which probably correlates with remobilization. Minerals indicative of the amphibolite facies (Turner, 1968) include almandine, cordierite, potassium feldspar, and hornblende. Concurrent amphibolite facies overprinting of the granite-gneisses and high grade metasedimentary rocks is also confirmed by the presence of almandine, cordierite, andalusite, and hornblende, that contain mineral inclusions of orthopyroxene, green spinel, and sillimanite (Figs. 7A, B, C, D, E and F).

Other granitoid plutons, emplaced in both domains, show no evidence of an earlier granulite metamorphism. These bodies include the Arch Lake Granite, and Wylie Lake and Colin Lake granitic complexes, and many minor granitoid plutons (Figs. 8D, 8F). All these plutons are either of granite, granodiorite or quartz diorite composition and their metamorphic mineral assemblages indicate that amphibolite facies conditions prevailed during formation and emplacement.

Similar mineralogical relationships were noted in the Shield rocks exposed south of Lake Athabasca. However, the regional tectonic and structural relationships are less well defined because of limited outcrop. All metamorphic data from this region were obtained from granitoid rocks.

Emplacement of the granitoid bodies has resulted in the formation of mantled gneiss domes. This is especially evident in the Wylie Lake area where the Wylie Lake granitic complex is undergoing unroofing (Fig. 2) and in the Tulip Lake area where Slave Granite forms the core of an elliptical dome (Langenberg and Godfrey, 1977).

The northerly trend of the typically elongated plutonic masses indicates deformation with a compressional east-west axis, and stress release in a north-south direction at the time of emplacement. Different conditions persisted, however, in the area between Colin Lake and just south of Andrew Lake where east-west trending fold axes and widespread fold interference patterns in the gneisses of the superstructure (Godfrey and Peikert, 1963; Godfrey, 1964) indicate polyphase deformation.

Continued influence of the east-west compressional deformation is seen in the linear form of the Waugh Lake (5 km east of Andrew Lake) depositional basin (Watanabe, 1961) where a sequence of greywackes, siltstones, basic flows, and tuffaceous clastics accumulated towards the end of phase II. An unconformity is assumed to be at the base of this sequence because of the prolonged uplift and erosion that must have followed Archean deformation, and because of the contrast in metamorphic facies between Waugh Lake rocks and the adjacent older rocks. The Waugh Lake sedimentary rocks exhibit graded bedding (Fig. 8B) and the lavas have retained their trachytic texture.

There are no similar basins of deposition in the Shield in either the Leland Lake domain or in the area south of Lake Athabasca.

Phase III

Plastic deformation, high grade metamorphism and remobilization, most prominent in the geological history described thus far, were followed by a cataclastically dominated style of deformation (Fig. 8E). Insofar as the rocks show a more brittle response to imposed conditions of stress, it must be assumed that geothermal gradients generally were depressed, the likely result of extensive removal of cover rock. Impressive expressions of cataclastic modifications are represented in the development of major mylonite bands of the Allan and Warren Fault Zones (Godfrey, 1958) which pass through Charles and Leland Lakes respectively. The steeply dipping Allan Fault Zone attains a width

[1] Igneous rocks have been classified according to Streckeisen's (1967) system.

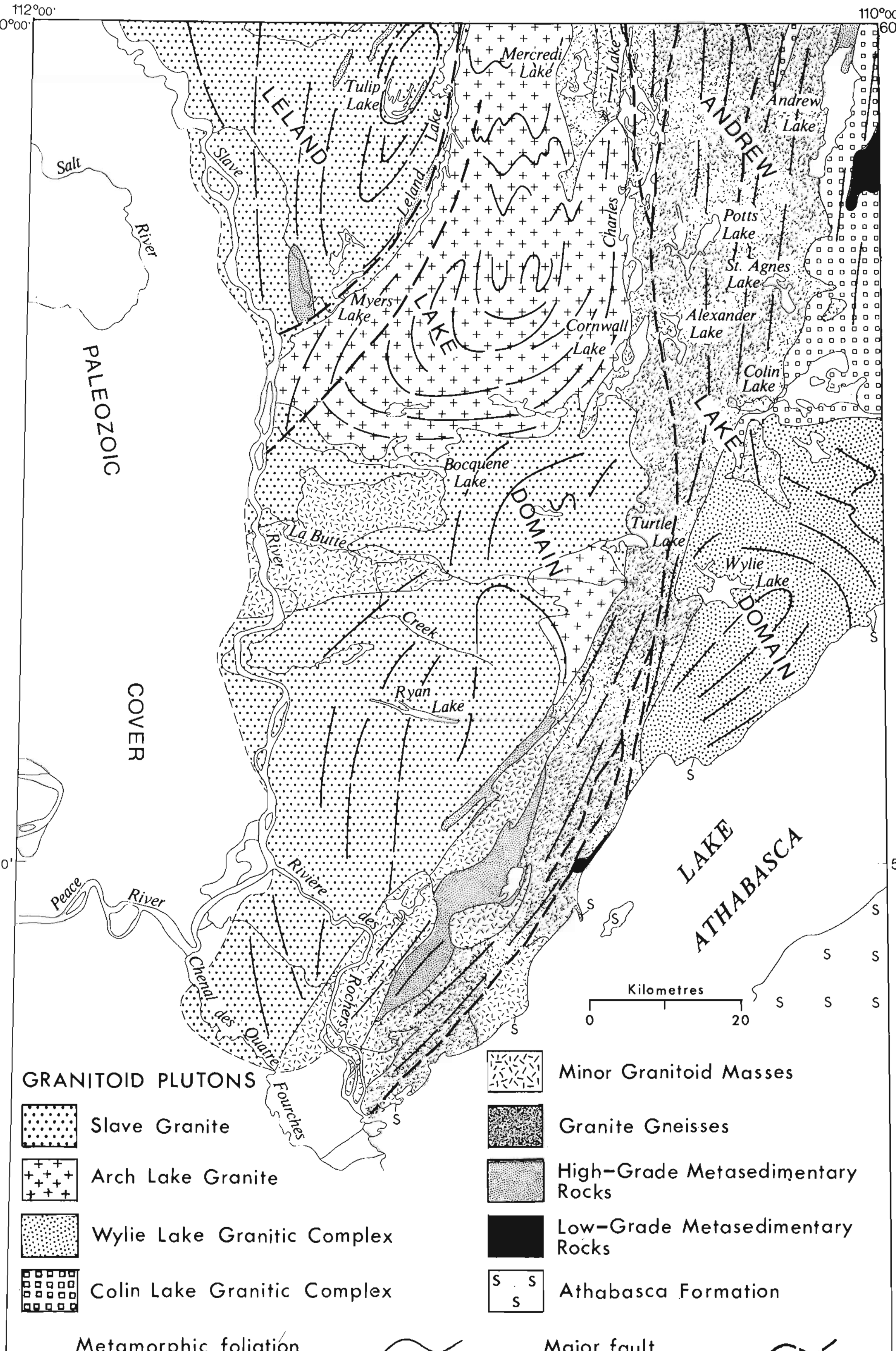


Figure 2. Bedrock geology map of the Canadian Shield, north of Lake Athabasca.

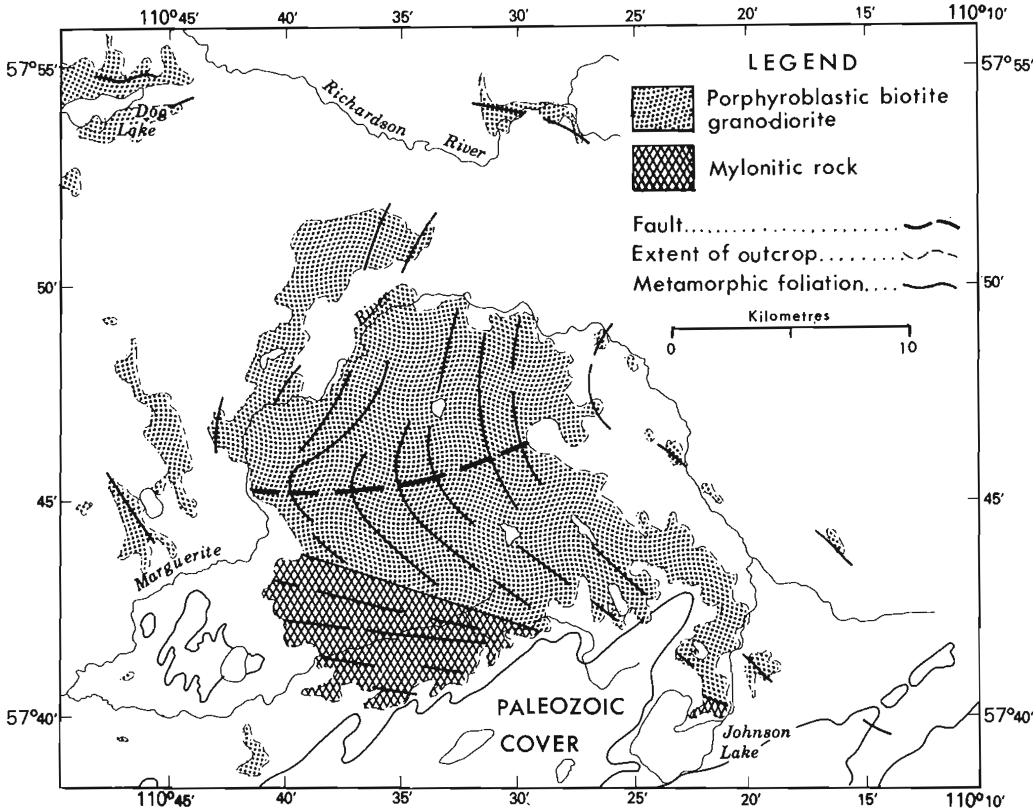


Figure 3.

Bedrock geology map of the Canadian Shield, south of Lake Athabasca.

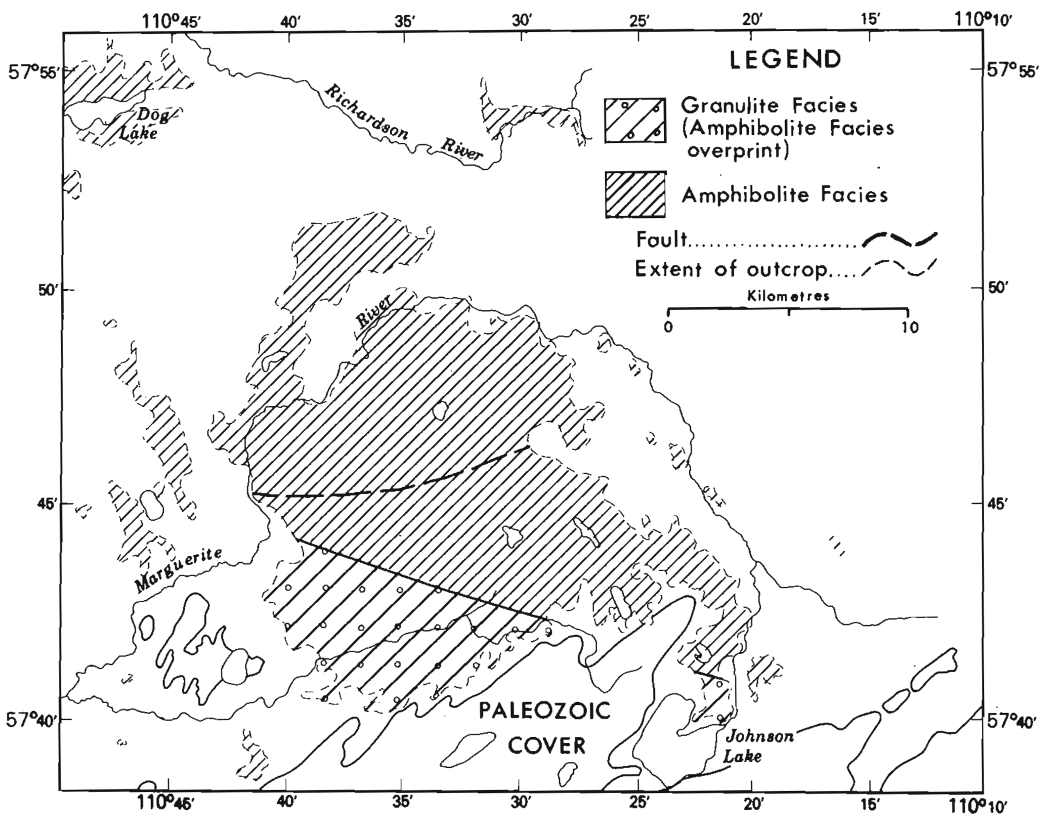


Figure 4.

Metamorphic facies map of the Canadian Shield, south of Lake Athabasca.

of up to 9.5 km in rocks which show a considerable range in degree of crushing in conjunction with extensive recrystallization. Rock textures reveal variable interaction between cataclasis and recrystallization, in which either process may have outlasted the other. Recrystallization took place under greenschist facies conditions. The principal cataclastic rock types in the mylonitic zones are ultramylonite, cataclasite, mylonite, and blastomylonite; less crushed sections show mortar structure, flaser gneiss, and porphyroclastic structure (Watanabe, 1965).

Other expressions of cataclasis are seen in the fabrics of most major rock types in both the Leland Lake and Andrew Lake domains. Some of the granitoid plutons show a dynamically induced penetrative foliation over wide areas. One good example (Fig. 8F) is seen in the main body of the Arch Lake Granite bounded by the Allan and Warren Fault Zones. In the Slave Granite, cordierite and garnet porphyroblasts are commonly elongated to axial ratios of 1 to 10 (Fig. 6E), and sillimanite is commonly concentrated in parallel aligned crush-zones (Fig. 6F).

Cataclastic and dynamic metamorphic effects are also evident in the Waugh Lake metasedimentary rocks where phyllonites, phyllites, and slates have been derived from greywackes and siltstones. The metamorphism of these rocks took place under greenschist facies conditions, reflected in the widespread distribution of chlorite. Furthermore, it is apparent from the ubiquitous presence of chlorite and epidote that all of the Shield exposed in northeastern Alberta, north and south of Lake Athabasca, was affected by this late phase of metamorphism.

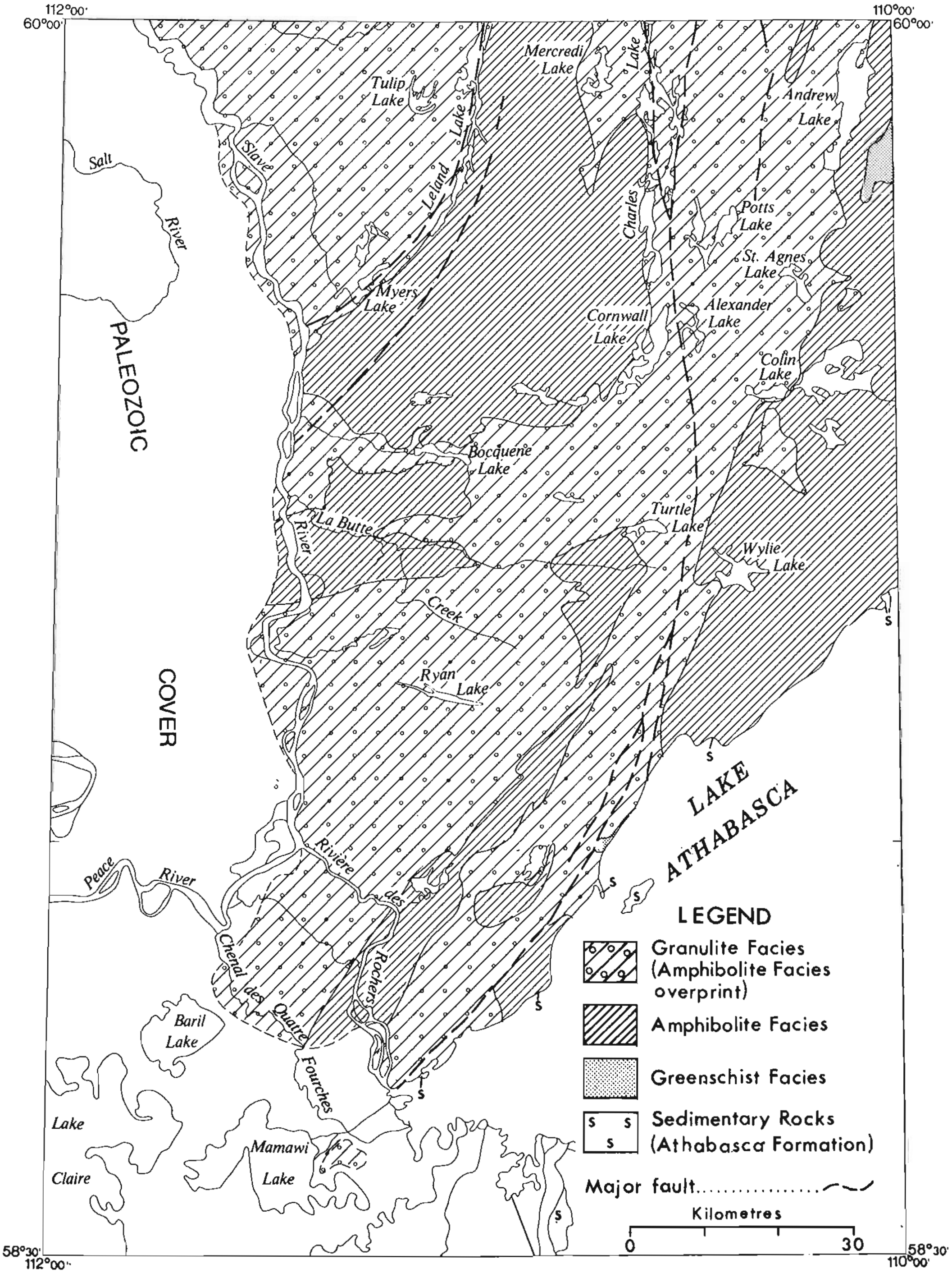
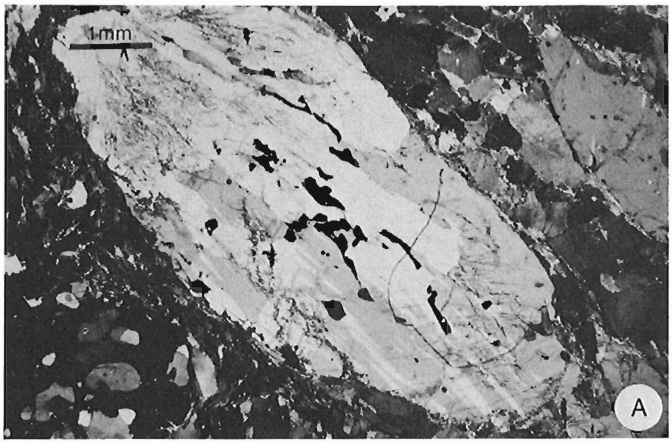
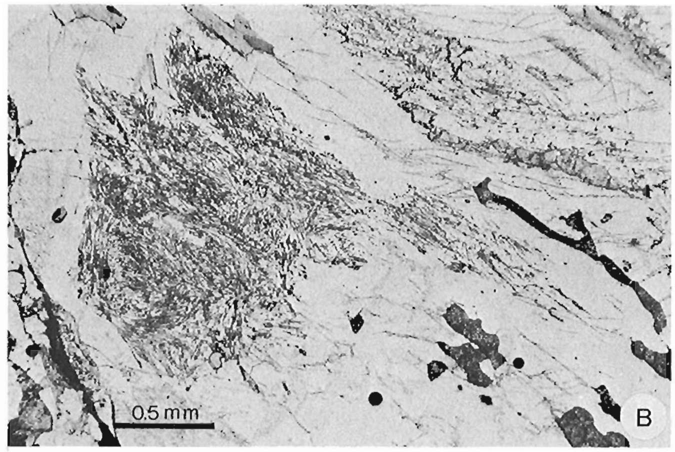


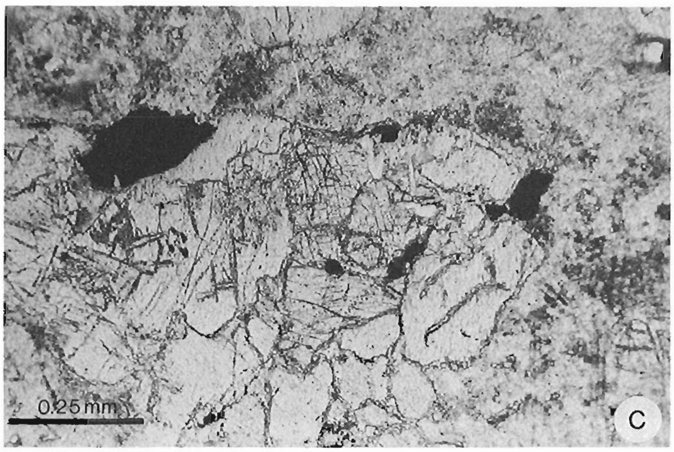
Figure 5. Metamorphic facies map of the Canadian Shield, north of Lake Athabasca.



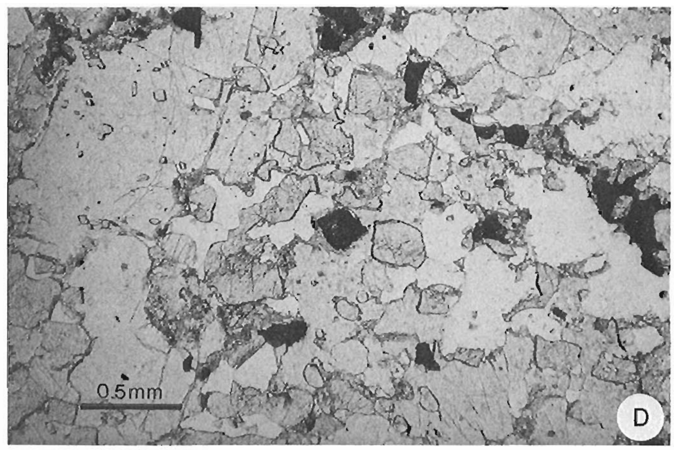
6A. Zone mineral relationships with a core of green spinel (dark) surrounded by sillimanite (white) in turn enclosed by biotite and almandine in Slave Granite. Plain light. (GSC 203309-A)



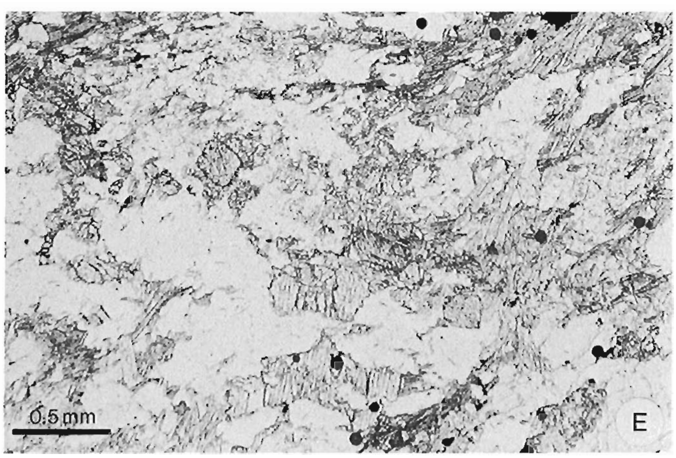
6B. Cordierite porphyroblast with characteristic twinning and pinite alteration in Slave Granite. Crossed nicols. (GSC 203309)



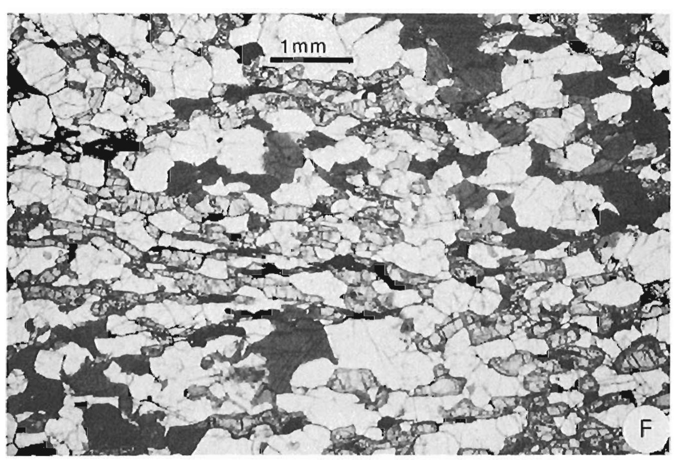
6C. Orthopyroxene, partly replaced by biotite along rim. Plain light. (GSC 203309-D)



6D. Late formed perthitic potassium feldspar porphyroblast with inclusions of plagioclase and quartz in Arch Lake Granite. Crossed nicols. (GSC 203309-B)

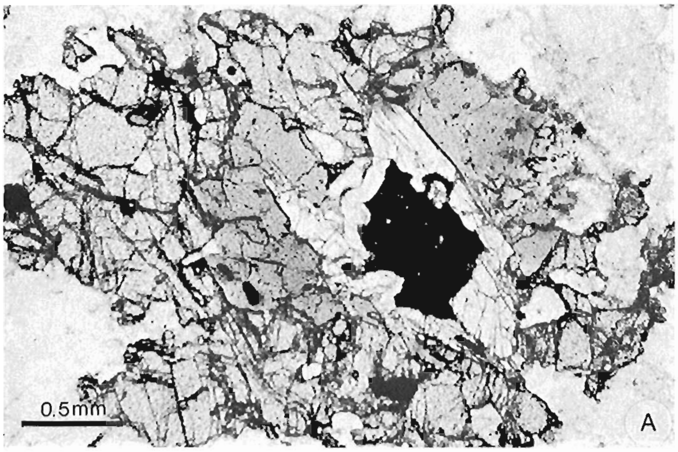


6E. Flattened garnet almandine crystal in Slave Granite, with parallel aligned green spinel (dark) fragments. Plain light. (GSC 203304-E)

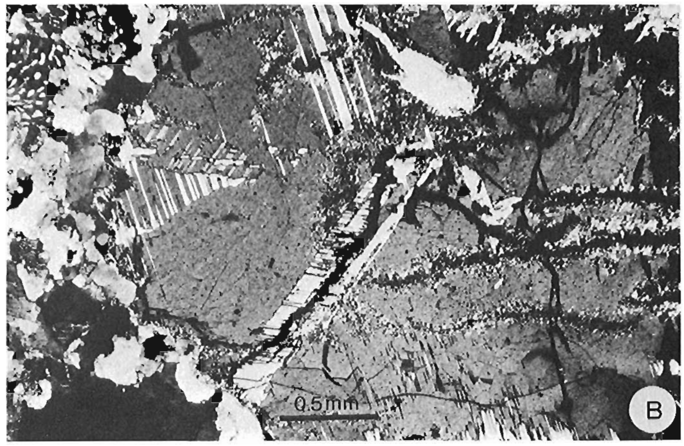


6F. Sillimanite concentrated in crush-zone between perthitic potassium feldspar porphyroblasts. Slave Granite. Plain light. (GSC 203309-D)

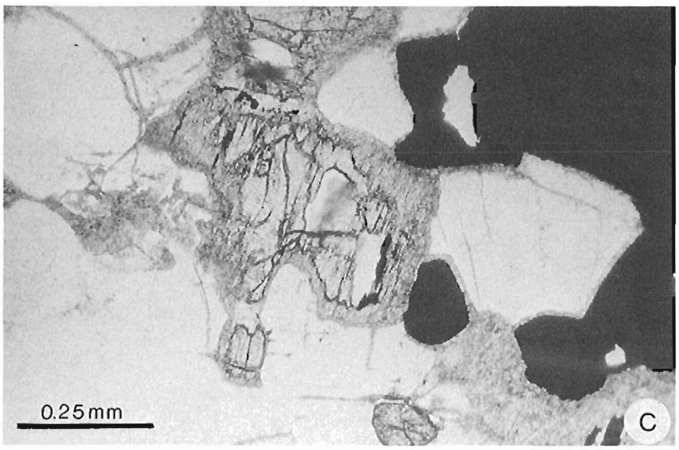
Figure 6. Photomicrographs showing metamorphic mineral relationships in granitoid rocks.



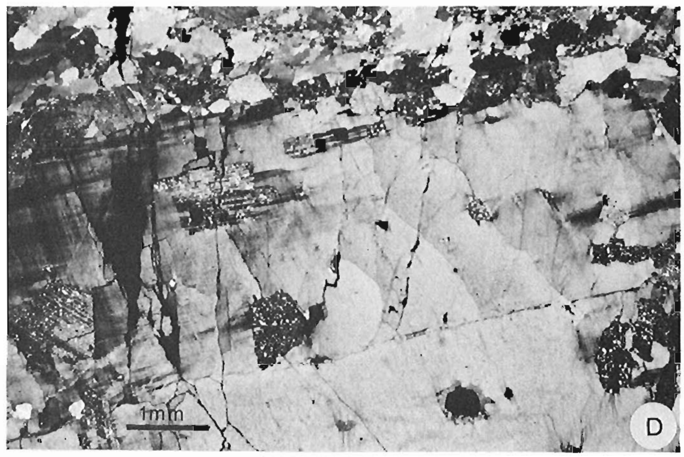
7A. Flattened cordierite porphyroblast enclosing green spinel (dark) and sillimanite (fine needles). Crossed nicols. (GSC 203309-G)



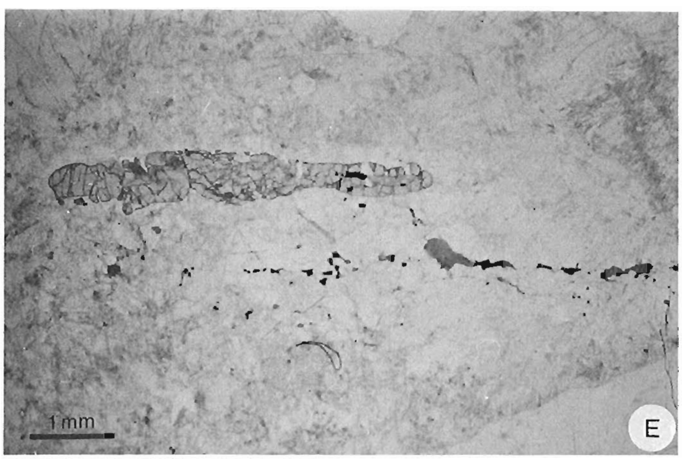
7B. Detail of cordierite porphyroblast in A, showing matte of sillimanite needles and dark patches of spinel. Plain light. (GSC 203309-F)



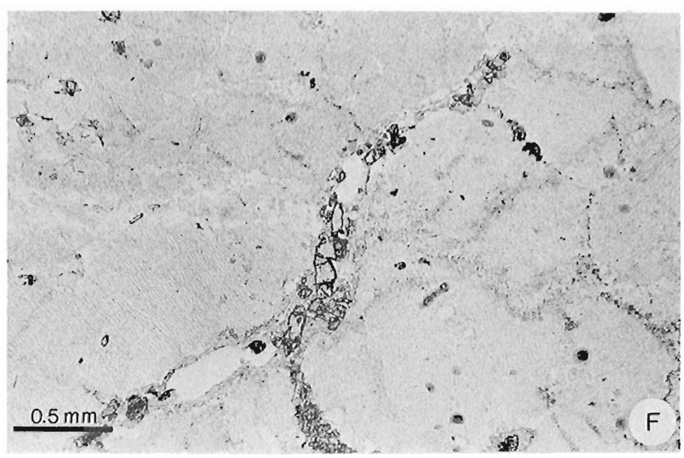
7C. Orthopyroxene (high relief, 90° cleavage) enclosed by light green amphibole (lower relief). Plain light. (GSC 203309-I)



7D. Sillimanite prisms (high relief), spinel (dark) and quartz (white) all enclosed by cordierite (grey). Plain light. (GSC 203309-H)



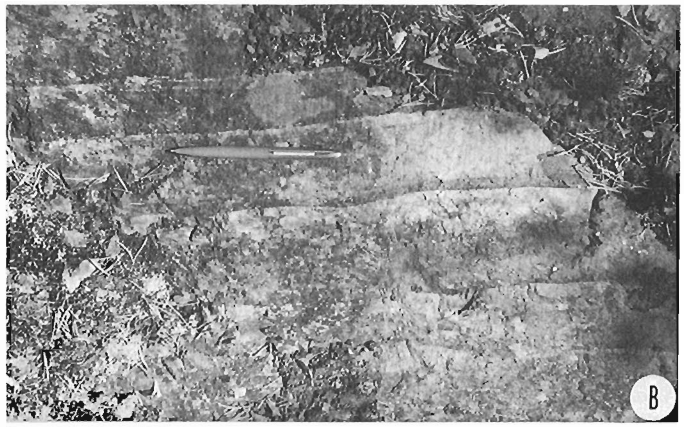
7E. Sillimanite prisms (rectangular intersection with 010 cleavage; high relief) enclosed by andalusite (lower relief). Plain light. (GSC 203309-K)



7F. Orthopyroxene (high relief) with preferred orientation, partly replaced by dark green amphibole without preferred orientation. Plain light. (GSC 203309-K)

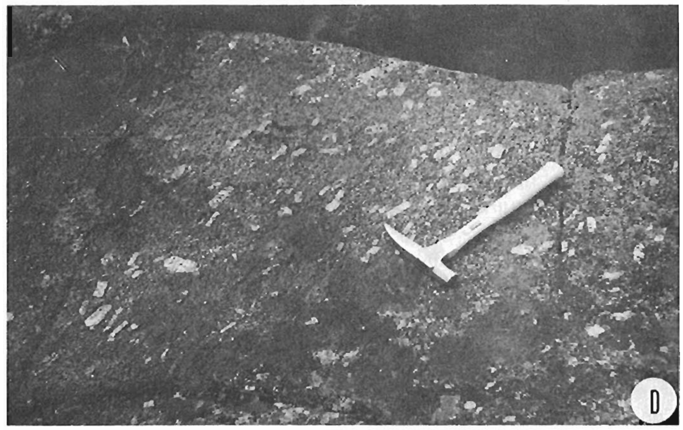
Figure 7. Photomicrographs showing metamorphic mineral relationships in metasedimentary rocks.

8A. High grade metasedimentary rocks in Andrew Lake domain; showing typical mineral banding, wavy folds and shears. (GSC 203309-N)

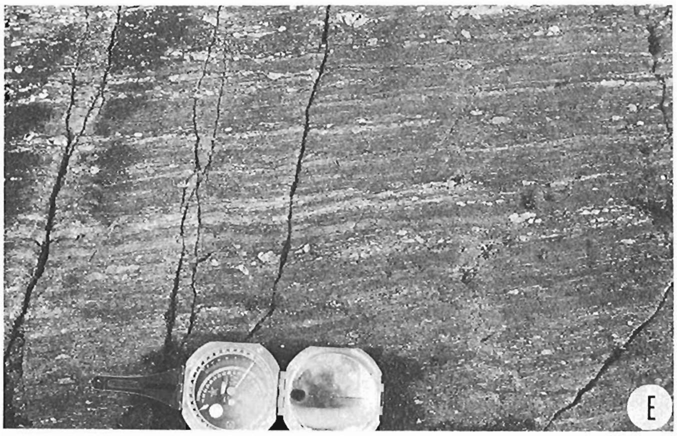


8B. Low grade metasedimentary rocks in Waugh Lake band; textural variation in graded beds is distinct. Slaty cleavage is oriented approximately perpendicular to bedding. (GSC 203309-L)

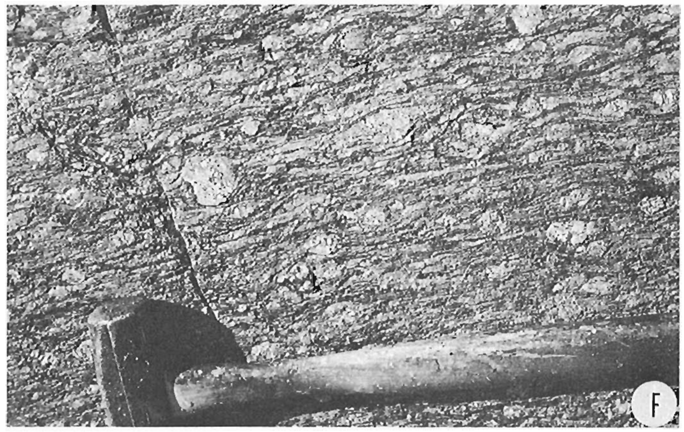
8C. Granite-gneiss terrane east of Charles Lake; migmatitic phase within hornblendic gneisses. (GSC 203309-O)



8D. Porphyroblastic Colin Lake granitic complex; showing typical preferred orientation of porphyroblastic potassium feldspars. (GSC 203309-N)



8E. Cataclastic gneiss, northeast of Fort Chipewyan; showing well banded structure with prophyroclastic feldspars in a mylonitic matrix. (GSC 203309-Q)



8F. Arch Lake Granite illustrating cataclastic nature of texture; with augen potassium feldspar and crushed streaky matrix aligned within a distinct foliation. (GSC 203309-P)

Figure 8. Outcrop photographs.

Deformation related to northerly fold axes, followed by mylonitization, Waugh Lake deposition, and greenschist facies metamorphism, essentially terminated the Hudsonian event in northeastern Alberta. A K-Ar biotite date of 1780 Ma from the Waugh Lake lavas effectively identifies this period in the geological history of the area (Baadsgaard and Godfrey, 1972).

Uplift and erosion resulted in stripping of the supracrustal granite-gneisses to varied degrees. Stripping in the southern part of the Andrew Lake domain has exposed the underlying Wylie Lake granitic complex. The Leland Lake domain has experienced a similar but more extensive erosion, with the result that the only remaining mantling gneisses are at the extreme north edge of the study area (Fig. 2). Erosion may have cut to a deeper crustal level on the west side of the Allan Fault Zone, indicating differential vertical displacement on the fault. Preferential removal of mantling gneisses in the southern part of both domains suggests a possible hinged uplift.

During continued uplift and erosion, continental conditions of sedimentation prevailed, and sandstones of the Helikian Athabasca Formation now lie unconformably on the older crystalline basement rocks.

SYNTHESIS AND SUMMARY

The Canadian Shield of northeastern Alberta is interpreted as part of a mobile zone which initially developed under granulite facies conditions during the Archean. The regional structural elements of the zone have a predominant northerly to northeasterly trend. Throughout the Aphebian (and late Archean?), remobilized infrastructure and reconstituted supracrustal rocks (in part) played a role in the emplacement of extensive granitoid plutons in the existing superstructure under amphibolite facies conditions.

Late Hudsonian deformation is considered to be responsible for development of the major north-south mylonite belts (mainly in the Andrew Lake domain), prominent cataclastic effects within the granitoid plutons (for example, the Arch Lake Granite) and a regional retrograde greenschist facies metamorphism. The effects of this metamorphism are best preserved in the lithologically and structurally distinct geosynclinal deposits of the Waugh Lake area.

Uplift in the area north of Lake Athabasca has resulted in the unroofing of granitoid plutons (for example, the Wylie Lake granitic complex). The mantling roof gneisses remain continuous only in the northern part of the study area.

ACKNOWLEDGMENTS

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METAMORPHIC PATTERNS AND THEIR RELATION TO TECTONISM AND PLUTONISM IN THE CHURCHILL PROVINCE IN NORTHERN SASKATCHEWAN

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Lewry, J.F., Sibbald, T.I.I., and Rees, C.J., Metamorphic patterns and thier relation to tectonism and plutonism in the Churchill Province in Northern Saskatchewan; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 139-154, 1978.

Abstract

Key metamorphic indicators, mineral assemblages and deformation/recrystallisation age relations have been collated to produce a metamorphic map of the Canadian Shield in Saskatchewan. Reactions defining facies boundaries and those allowing more detailed facies subdivision are discussed.

The pattern thus defined shows broad areas of upper amphibolite to granulite facies metamorphism (low pressure-intermediate facies series) with less extensive, and commonly rather narrow, intervening zones of upper greenschist to lower amphibolite facies parageneses. A close spatial relation to major lithostructural domains and crustal discontinuities is observed. In the west particularly, but also elsewhere, steepening of isograd surfaces and transition to low grade zones is related to major shear zones with a long history of repeated movement. Such relations are examined and a major junction between lightly reworked, polymetamorphic Archean craton and thoroughly remobilized, essentially monometamorphic Hudsonian mobile belt is defined.

Within the Hudsonian mobile belt the broad high grade metamorphic zones of extensively migmatized supracrustals and remobilized infrastructure are characterized by a high proportion of in situ anatectic mobilizate but generally lack significant development of discrete allochthonous Hudsonian plutons. It is suggested that within these zones early tectonic upward and outward flowage of remobilized infrastructure was a major factor in crustal heat transfer and establishment of later thermal patterns. The lower grade areas, where not related to major shear zones, are broadly coincident with development of an extensive sequence of early to late tectonic Hudsonian plutons, a situation exemplified by the central and eastern parts of the province. Migmatization in such areas is largely a magmatic injection phenomenon.

The thermotectonic domains thus recognizable, representing differing elements of the Hudsonian Orogen, are described and their significance and interrelationships briefly discussed. Available evidence suggests that thermal patterns and isograds underwent little migration during tectonism throughout most of the belt and that generally only one broad thermal peak or plateau was achieved, punctuated by successive deformational events which triggered several episodes of metamorphic recrystallization and mineral growth.

Some data indicating lower pressures towards the termination of thermotectonism (possibly by stripping of cover?) are discussed. There is no clear metamorphic expression of major crustal suture/subduction within this part of the Hudsonian belt of the Churchill Province.

Résumé

On a comparé les assemblages minéraux et les corrélations entre les âges de déformation et de recristallisation, pour produire une carte métamorphique du Bouclier précambrien de la Saskatchewan. Dans le présent travail, on étudie les réactions qui permettent respectivement de définir les limites de faciès, et d'effectuer une subdivision plus détaillée des faciès.

On a pu ainsi définir de vastes ensembles métamorphiques, qui appartiennent aux faciès des amphibolites supérieures et des granulites (suite de faciès de pression faible et de température intermédiaire), ainsi que des zones intermédiaires moins étendues et plutôt étroites caractérisées par des paragenèses allant du faciès des schistes verts supérieurs à celui des amphibolites inférieures. On a observé une étroite corrélation spatiale entre ces éléments, et les principaux domaines lithostructuraux ainsi que les discontinuités de la croûte terrestre. A l'ouest en particulier, mais aussi ailleurs, la grande inclinaison des surfaces d'isogrades et la transition soudaine vers des zones de plus faible degré métamorphique sont liées à de grandes zones de cisaillement, qui ont fréquemment subi des mouvements. Dans le présent article, on examine ces relations, et on définit une jonction importante entre le craton archéen polymétamorphique, légèrement remanié, et la Zone mobile hudsonienne complètement remobilisée, et surtout monométamorphique.

Dans la Zone mobile hudsonienne, les vastes zones qui ont été soumises à un métamorphisme de degré élevé et sont composées de roches supracrustales migmatisées sur une grande étendue et fortement remobilisées, sont caractérisées par une proportion élevée de mobilisat anatectique formé in situ, et on y trouve généralement peu de plutons hudsoniens allochthones discrets. On suggère que dans ces zones, un écoulement tectonique ancien ascendant et dirigé vers l'extérieur de l'infrastructure soumise à une remobilisation a joué un rôle important dans le transfert thermique qui a eu lieu à l'intérieur de la croûte terrestre, et dans l'établissement du régime thermique ultérieur.

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Les zones de degré métamorphique inférieur, lorsqu'elles ne sont pas liées à des zones de cisaillement importantes, coincident approximativement avec le développement d'une succession étendue de plutons hudsoniens tectoniques, anciens ou plus récents, situation que l'on rencontre dans les parties centrale et est de la province. Dans ces zones, la migmatisation est principalement un phénomène d'injection magmatique.

On décrit ici les domaines thermotectoniques ainsi identifiables, qui représentent les divers éléments de l'orogène hudsonien, et l'on comment brièvement leur importance et la corrélation qui existe entre eux. Les indices dont on dispose semblent indiquer que les régimes thermiques et les isogrades n'ont subi qu'une faible migration au cours du tectonisme dans la majeure partie de la zone, et que généralement, il n'y a pas eu plus d'un maximum ou d'un plateau thermiques qui ont été marqués par des épisodes successifs de déformation, lesquels ont provoqué plusieurs épisodes de recristallisation métamorphique et de développement des minéraux. Certaines données qui indiquent un abaissement des pressions vers la fin du thermotectonisme (peut-être par disparition de la couverture rocheuse?) sont examinées. On observe pas d'expression métamorphique définie d'une importante zone de subduction ou d'une importante suture de la croûte terrestre, dans cette portion de la zone hudsonienne de la province de Churchill.

INTRODUCTION

This paper presents the results of an ongoing compilation and analysis of metamorphic data from the Canadian Shield in northern Saskatchewan. The general spatial distribution of key metamorphic indicators and metamorphic facies is described and the relationship between this distribution and the major lithological and structural subdivisions within the region is considered.

Data used in preparation of a metamorphic map of the northern Shield area have been derived mainly from work published since 1949 by the Saskatchewan Geological Survey. North of 58°N mapping by the Saskatchewan Geological Survey has been confined to the northwest corner of the province, the area north of Stony Rapids, the eastern boundary of the province, and the Wollaston Lake area. In this region metamorphic data have also been obtained from maps and reports published by the Geological Survey of Canada. The compilation is supported by some relatively detailed metamorphic studies, which are summarized in this report.

Metamorphic mineral data were collated by one of the authors (C.J.R.), as overlays on an updated geological map of northern Saskatchewan. Occurrences of andalusite, sillimanite, kyanite, staurolite, cordierite, almandine, muscovite, hypersthene, and diopside were plotted and, where available, plagioclase and amphibole compositions were recorded. The degree of anatexis was also noted and, where possible, distinction has been made between the processes of in situ and magmatic injection migmatization.

Presently available data do not provide a complete or consistent coverage of the region and, in particular, there is a general lack of information on age relations between mineral growth, deformation, and plutonism. However, most regions of significant polymetamorphic overprinting appear to have been identified.

Description and discussion of the results of the metamorphic compilation are preceded by a summary of the lithological and structural framework of the Canadian Shield in Saskatchewan. The spatial and temporal patterns of metamorphism are then discussed in the context of regional crustal evolution.

GENERAL GEOLOGY

Recent work (Lewry, 1974; Sibbald, 1974; Munday, 1974; Ray, 1974, 1975; Gilboy, 1975; Lewry and Sibbald (1977), Sibbald et al., 1977) has shown that the southern part of the Canadian Shield of northern Saskatchewan is divisible into a number of discrete crustal units or lithostructural domains, each possessing general internal uniformity of lithology and structure. A modification of the lithostructural subdivision proposed by Lewry and Sibbald (1977), involving some grouping of previously defined domains and extrapolation into the northern part of the province, forms the basis of the summary presented below.

From west to east four broad crustal units were recognized (Fig. 1): Western Craton, Cree Lake Zone, Rottenstone Complex, and Southeastern Complex.

Western Craton

The Virgin River Shear Zone (Fig. 1), a narrow belt of blastomylonitic gneiss and foliated, late tectonic, granitic intrusives lying to the south of the Athabasca Formation, is now recognized as a major crustal discontinuity between an Archean craton to the west and a Hudsonian mobile zone to the east (Lewry and Sibbald, 1977). On the basis of general lithological, structural, and geophysical evidence, this major locus of repeated ductile to brittle shearing is thought to be represented north of the Athabasca Formation by the Black Lake Fault Zone.

In the southwest, two Archean crustal units west of the shear zone (the Western Granulite domain and the Firebag domain) comprise granodioritic gneiss, subordinate metagabbro, anorthosite, and minor supracrustal remnants. These have been only lightly reworked by Hudsonian thermotectonic events, except within the Virgin River Shear Zone. The two units are separated by the Clearwater domain, interpreted as a narrow subsidiary Hudsonian mobile zone, which is largely obscured by Phanerozoic cover (Fig. 1).

North of the Athabasca Formation most crystalline rocks lying to the west of the Black Lake Fault Zone have been collectively termed the Tazin Group (Tremblay, 1968). On the grounds of radiometric dating and lithostructural similarities they are considered to form the northern extension of the western craton. Linear zones of intense Hudsonian reworking are recognizable in the craton (Beck, 1969) and aeromagnetic patterns indicate that one of these, the Grease River Belt, may be continuous with the Clearwater domain to the south (Wallis, 1970a). In contrast to the southern part of the craton, it is probable that some Aphebian supracrustal cover rocks are also preserved (Krupička and Sassano, 1972; Morton and Sassano, 1972; Baadsgaard and Godfrey, 1967; Koster and Baadsgaard, 1970; Scott, 1972).

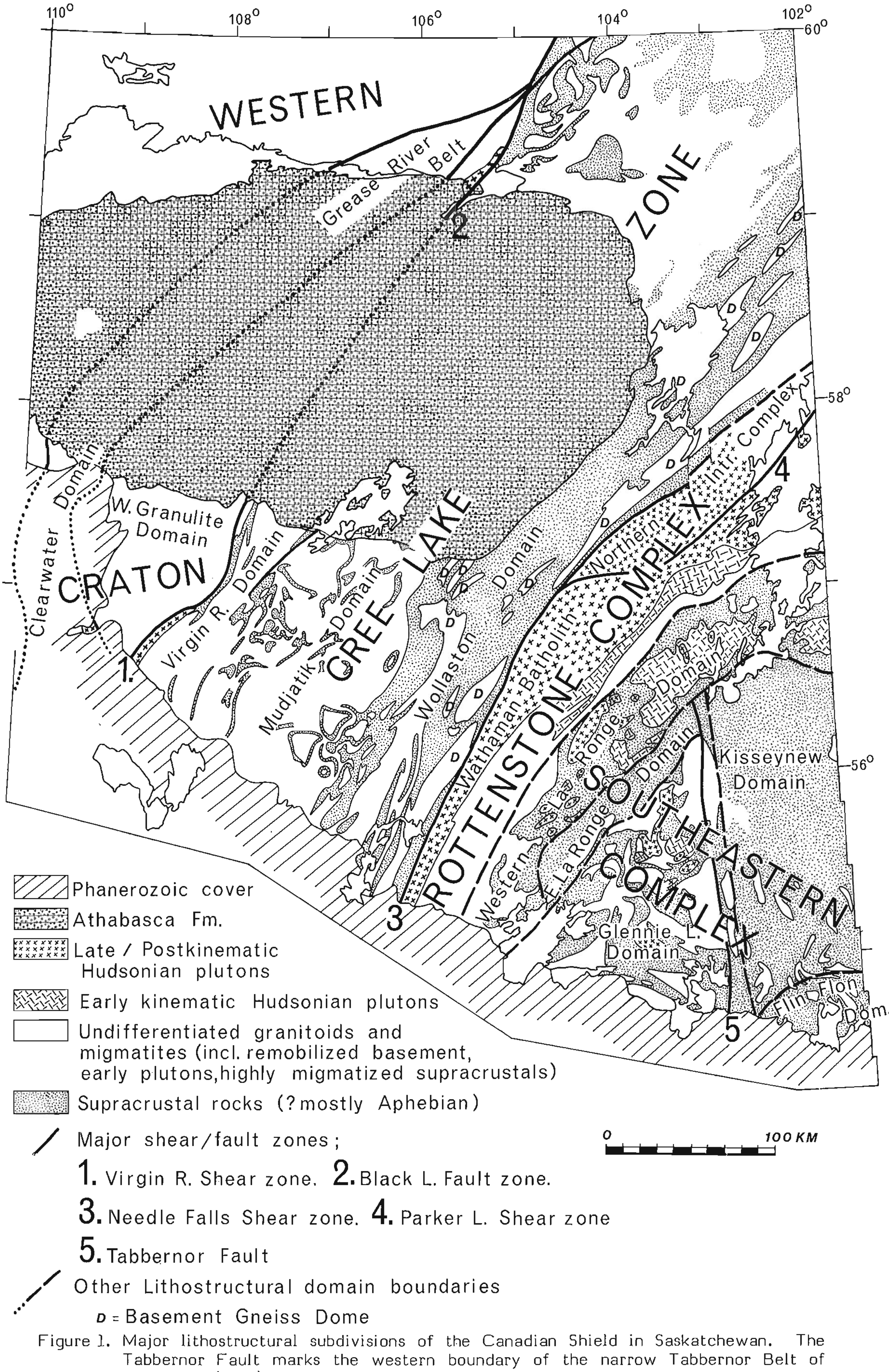

Figure 1. Major lithostructural subdivisions of the Canadian Shield in Saskatchewan. The Tabbernor Fault marks the western boundary of the narrow Tabbernor Belt of Macdonald (1975).

Cree Lake Zone

To the south of the Athabasca Basin the Cree Lake Zone is bounded on the west by the Virgin River Shear Zone and on the east by the Needle Falls Shear Zone. The Hudsonian evolution of the Creek Lake Zone is considered to have proceeded by remobilization of the Archean basement (Lewry and Sibbald, 1977, Sibbald et al., 1977). In the more mobile central region (Mudjatik domain) the plastic infrastructure, which probably incorporated basal portions of the Aphebian cover, moved upwards and outwards as flat lying migmatite nappe lobes. The overlying, less modified, Aphebian supracrustals were repeated in constricted synclinal belts at different structural levels. At the cooler margins of this zone, in the western parts of the Virgin River domain and in the Wollaston domain, less mobile gneiss domes developed rather than nappes.

Later phases of compressive strain refolded the flat lying central nappe core along steeply dipping northwesterly (D_2) and northeasterly (D_3) trending axial surfaces to yield a dome and basin outcrop pattern (Sibbald, 1973, 1974; Lewry, 1974; Pearson and Lewry, 1974). In the marginal zones, however, pre-existing gneiss domes were flattened into doubly plunging, northeasterly elongated structures. Complete structural and lithological transition occurs between the different domains. Hudsonian granitoids in the Cree Lake Zone are largely of in situ anatectic origin.

To north and east of the Athabasca Formation the Cree Lake Zone extends from the Black Lake Fault Zone eastwards to the southeastern boundary of the Wollaston domain. Closed structures and arcuate Aphebian supracrustal belts due east of the Black Lake Fault Zone, in the Charlebois Lake area (Mawdsley, 1957), are interpreted as mantled gneiss domes marginal to a more mobile core. From structural relations and radiometric dating in the adjacent Kasmere Lake area of northwestern Manitoba (Weber et al., 1975) the granitoid inliers northeast of Wollaston Lake have been interpreted as a northeasterly extension of the flattened gneiss dome regime of the Wollaston domain. The intervening region is poorly documented and its structure unknown; it is here considered to be an extension of the nappe core of the Cree Lake Zone, possibly exposed at a higher tectonic level than to the south. The southeastern boundary of the Cree Lake Zone is mainly defined by the Needle Falls Shear Zone, but its northeasterly extension is imprecisely located.

Rottenstone Complex

The Rottenstone Complex (Lewry and Sibbald, 1977) comprises a broad belt of plutonic rocks of presumed Hudsonian age, and subordinate migmatized supracrustal rocks whose neosome is largely of magmatic injection origin (Gilboy, 1975; Munday, 1974; Lewry, 1975, 1976; Ray, 1974, 1975; Stauffer et al., 1976). Early syntectonic, dioritic, tonalitic, and granodioritic schleiric gneiss as well as migmatite are intruded by late- to post-tectonic plutons of monzonitic to granitic composition, represented mainly by the Wathaman Batholith.

The eastern part of the Rottenstone Complex is formed by a belt of multistage migmatite, which pass eastwards into essentially unmigmatized rocks of the western La Ronge domain. In the northeast a distinct crustal subunit, the Northern Intrusive Complex, comprising abundant diorite-gabbro intrusives and later granite, is separated from the Wathaman Batholith by the Parker Lake Shear Zone (Fig. 1).

Southeastern Complex

The Southeastern Complex is poorly understood; however, it is divisible into a number of lithostructural domains (Fig. 1; and Lewry and Sibbald, 1977).

The western La Ronge domain, lying directly east of the Rottenstone domain, is a belt of generally unmigmatized supracrustal rocks invaded by high level plutons, of various ages. The supracrustal rocks are predominantly basic, with subordinate acid metavolcanics and volcanogenic metasediments. The plutons range in composition from dioritic to granitic, and in age from early to post-tectonic; they dominate the northern part of the western La Ronge domain (Lewry, 1975). Although the form and probable level of emplacement of the plutons is dissimilar to that in the Rottenstone domain, the compositional range and age sequence of plutonism is essentially the same. The western La Ronge domain is separated from the Rottenstone domain by an intervening injection migmatite belt.

The eastern part of the La Ronge domain differs markedly from the western part; it possesses a more prominent structural linearity and lacks ovoidal plutons. Early syntectonic elongate gneissic granite are subordinate to an assemblage of predominantly pelitic to semipelitic gneiss. Minor psammitic, calc-silicate, and mafic gneiss also occur and the supracrustals as a whole appear similar to the paleosome fraction of the migmatites of the eastern Rottenstone domain. This gneiss has been correlated (Kirkland, 1959; Johnston, 1968) with the type Kisseynew gneiss of Bruce (1918) and the northeastern part of the eastern La Ronge domain appears to be continuous with the main Kisseynew domain (see below). The contrast in lithological character and structural/intrusive style between the western and eastern parts of the La Ronge domain is emphasized by the occurrence of a discontinuous meta-arkosic unit at their common boundary. This arkosic unit may correlate with the Sickle Group in Manitoba.

On the east the La Ronge domain passes transitionally into the Glennie Lake domain, which is characterized by nonlinear, arcuate to closed outcrop and foliation patterns (Lewry, 1978; and Fig. 1). It is formed by an intrusive sequence of granitoid rocks ranging from early dioritic-granodioritic gneiss, more homogeneous syntectonic granodiorite-tonalite bodies to late and post-tectonic leucogranodiorite and alaskite plutons. Subordinate gabbroic plutons are also present. The supracrustal rocks occur as narrow arcuate septa and isolated infolded belts between intrusions. The structural pattern is considered to result both from fold interference and sequential diapiric emplacement of the granitoid masses (Lewry, 1978). Supracrustal lithologies are similar to, and continuous with, those in adjacent domains.

The eastern margin of the Glennie Lake domain is defined by the Tabbernor Belt (Macdonald, 1975), a narrow northerly trending fold-fault zone. The Tabbernor Fault (sensu stricto), although the most prominent feature of the belt, is now recognized as only one element within a restricted zone of Hudsonian mylonites and related north-south trending upright folds (Sibbald, in press; Macdonald 1976). The combined "Belt" extends to the margin of the Rottenstone Complex. Farther north it manifests itself only as a series of subparallel late faults.

The Kisseynew domain is a region of high grade migmatitic supracrustal rocks (Kisseynew gneiss) and minor granitoids. The supracrustal rocks comprise a predominantly pelitic to semipelitic, commonly graphitic, greywacke assemblage correlated with the Burntwood River Supergroup of Manitoba, and subordinate mafic gneiss and psammites, in part correlated with the Wasekwan and Sickle groups respectively. These rocks have undergone, in some areas at least, recumbent nappe folding (Pearson 1972, Macdonald 1975; Gilboy, 1976), followed by refolding along near vertical north-south (D_2) and northeast-southwest (D_3) trending axial surfaces (Pearson, 1972). Granitoid bodies are generally gneissic, and in part may represent early nappe cores.

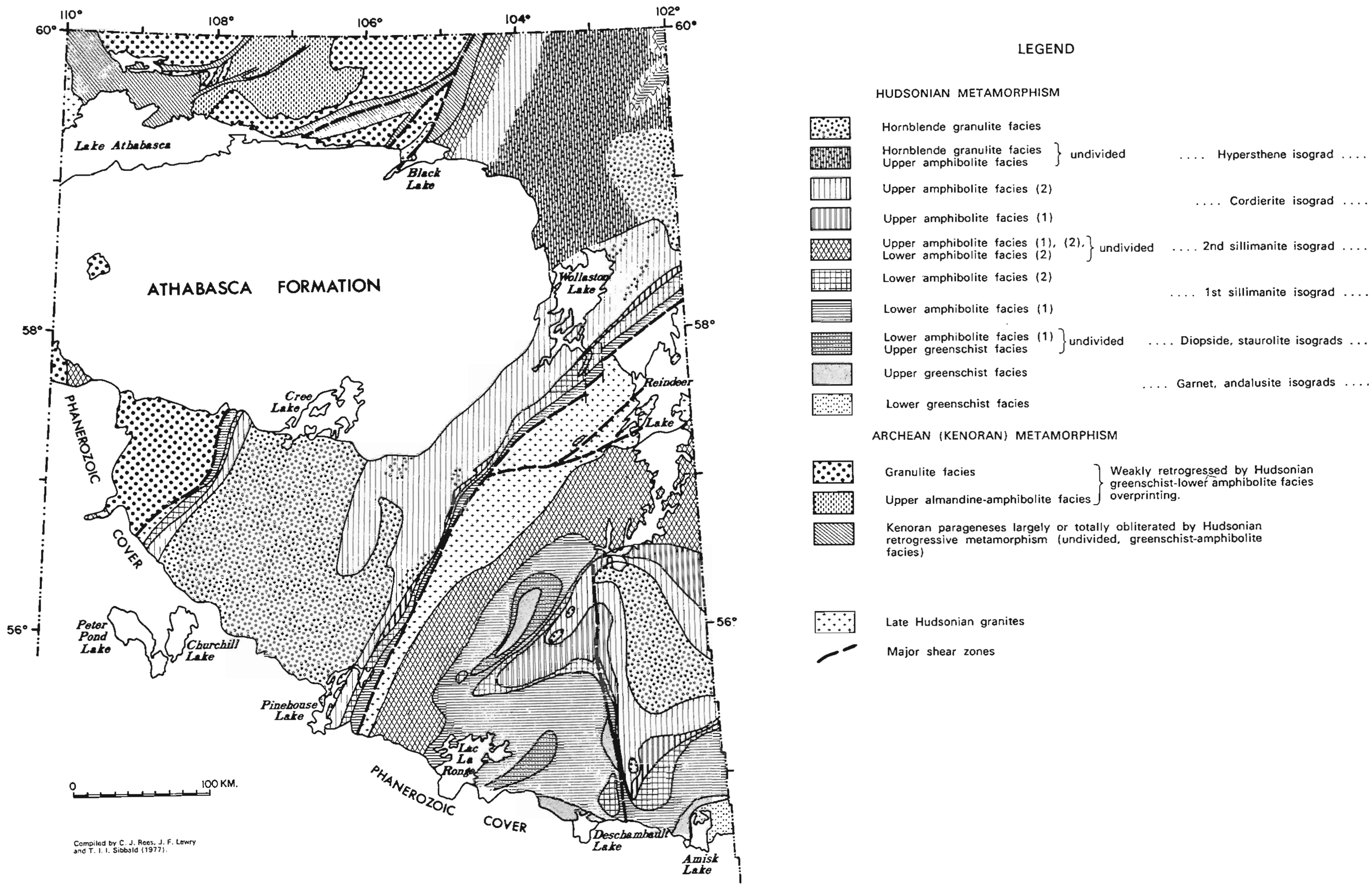

Figure 2. Metamorphic map of the Canadian Shield in Saskatchewan. Subdivisions of the amphibolite facies lower amphibolite facies (1) and (2) and upper amphibolite facies (1) and (2) are based on the reactions specified in the text.

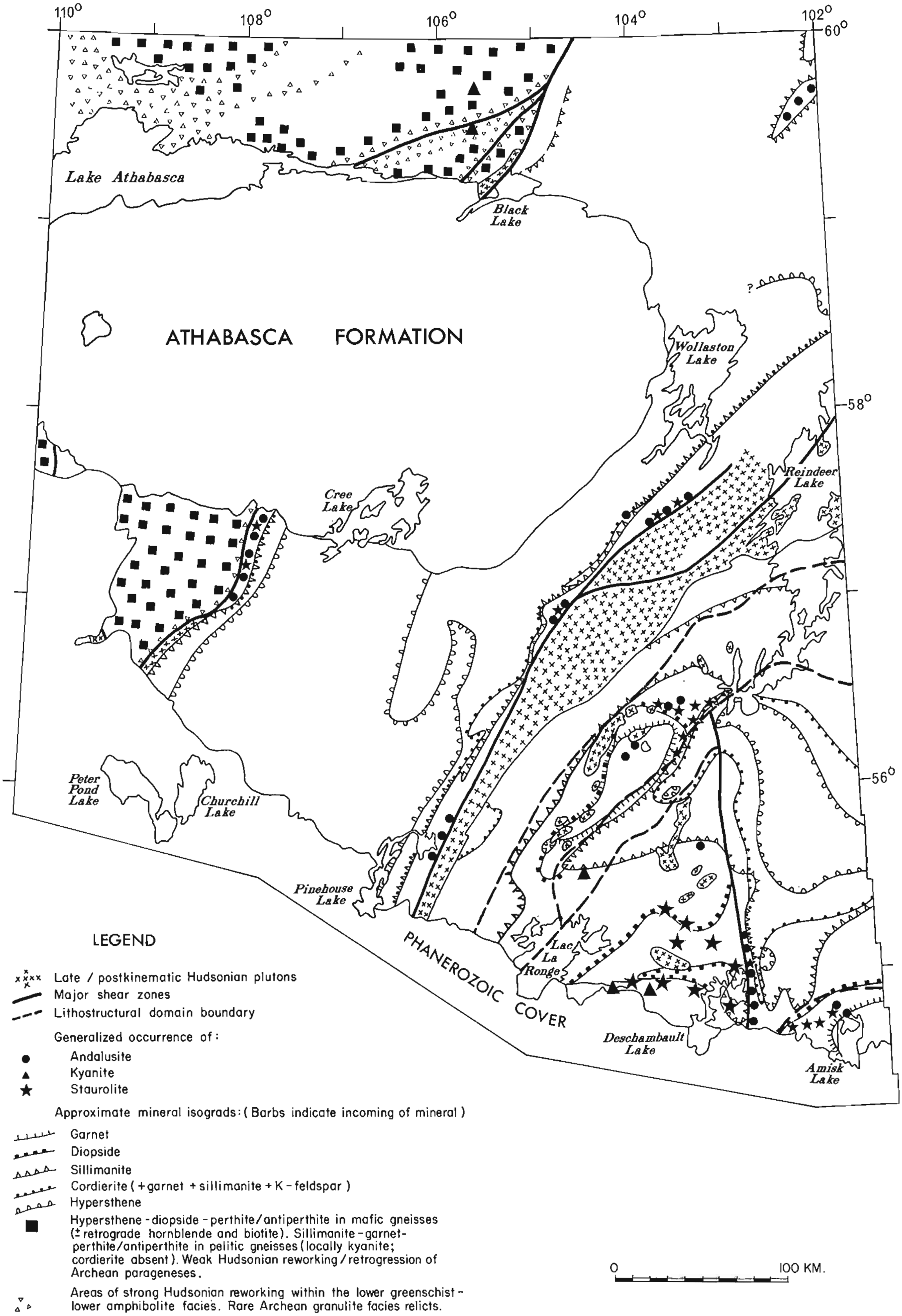

Figure 3. Metamorphic mineral distribution and isograds.

In the Flin-Flon domain lithologies typical of the Burntwood River Supergroup are transitional to metavolcanics (Amisk Group) and coarse clastic metasediments (Missi Group). Plutonic bodies of gabbroic to granitic composition, of syntectonic to post-tectonic age, are also present. As indicated by numerous early workers (Harrison, 1951; Byers and Dahlstrom, 1954), the junction between the domains, the 'Kisseynew front', is a metamorphic transition as well as a lithological and structural transition as shown by Stauffer and Mukherjee (1971) and Pearson (1972).

The supracrustal rocks within the Southeastern Complex are considered to be largely Aphebian (Mukherjee et al., 1971; Bell et al., 1975) although some Archean remnants are recorded (Coleman, 1970). The plutonic sequence is also mainly of Hudsonian age, but includes some Archean granitoid components (Coleman, 1970; Macdonald, pers. comm.). Early granodioritic gneiss such as that in the Glennie Lake domain may be in part remobilized basement. However, there is no conclusive evidence that the Aphebian supracrustals were deposited on Archean continental crust throughout this whole region, as is the case west of the Needle Falls Shear Zone.

METAMORPHIC SUBDIVISION

The metamorphic map (Fig. 2) is based primarily on the distribution of minerals and mineral assemblages in pelitic rocks (Fig. 3). In the absence of pelites, metamorphic assemblages in other rock types have been incorporated, as have generalizations relating to the degree and character of anatexis.

Thin section studies of relationships between deformation phases and mineral growth, together with field observations, have served to distinguish major areas of polymetamorphism, for example upper greenschist-lower amphibolite facies (Hudsonian) overprinting of the granulite facies (Archean) metamorphism in the Western Craton. In general, such studies have shown the temporal uniformity of Hudsonian metamorphism, a feature which contrasts with the episodic, variable grade character of metamorphism in some other orogenic belts; for example, the Western Alps (Chatterjee, 1961; Bearth, 1958, 1962) and the Scottish Caledonides (Johnson, 1963; Rast, 1963). Thus the metamorphic facies map may illustrate the thermal pattern throughout Hudsonian time as well as that which existed during the last major period of recrystallization. The criteria used to define the subdivisions represented on the metamorphic map of Saskatchewan (Fig. 2) are summarized below.

I. Hudsonian Metamorphism

a) Greenschist facies

Two subdivisions of the greenschist facies are recognized. The higher grade is differentiated from the lower by the occurrence of garnet, presumed to be almandine (garnet isograd) and/or andalusite (andalusite isograd).

b) Greenschist-amphibolite facies transition

Minerals stable in the greenschist (and/or lower) facies are associated in this zone with minerals generally stable only in the amphibolite (and/or higher) facies.

c) Amphibolite facies

The appearance of minerals such as staurolite, cordierite, diopside, grossularite, and orthoamphiboles, and the disappearance of chlorite marks the commencement of the amphibolite facies. Within the amphibolite facies four subdivisions (lower amphibolite facies (1) and (2) and upper amphibolite facies (1) and (2)) are differentiated by: 1) the appearance of sillimanite and the disappearance of andalusite; 2) the appearance of orthoclase and sillimanite and the disappearance of muscovite; 3) the appearance of cordierite. These transformations have been ascribed to the following reactions:-

1st sillimanite isograd: andalusite = sillimanite (R.1)

2nd sillimanite isograd: muscovite + quartz = orthoclase + sillimanite + H_2O (R.2)

cordierite isograd: biotite + sillimanite = cordierite + almandine + K feldspar + H_2O (R.3)

d) Amphibolite-granulite facies transition

The appearance of hypersthene (hypersthene isograd) marks the attainment of the granulite facies. Hypersthene may appear in prograde sequences in mafic granulites, granitoid rocks, and pelitic metasediments.

The facies series which characterizes the Hudsonian thermal imprint can be broadly inferred from the observed distribution of critical minerals and mineral assemblages (Fig. 3). Upper pressure limits are set by the occurrence of andalusite and sillimanite and by a general absence of kyanite. Kyanite has been reported only in the Otter Lake and Wapawekka Lake areas (Padgham, 1963, 1966). A lower pressure limit, at the intersection of the univariant curves marking the first and second sillimanite isograds, is imposed by the generally observed prograde sequence andalusite + muscovite → sillimanite + muscovite → sillimanite + alkali feldspar.

Pelitic rocks in the lower amphibolite facies contain almandine(?) garnet and sparsely distributed cordierite; in the upper amphibolite and lower granulite facies, cordierite, associated with almandine and sillimanite, is widespread.

These occurrences indicate the existence of an Abukuma (low pressure), or a transitional Abukuma-Barrovian (low-medium pressure) facies series.

II. Archean (Kenoran) Metamorphism

Areas influenced by Kenoran metamorphism are subdivided as follows:

a) Granulite facies

b) Upper almandine-amphibolite facies

These assemblages are commonly weakly retrogressed by Hudsonian greenschist-lower amphibolite facies overprinting. Areas of strong retrogression in which Kenoran parageneses are largely obliterated by Hudsonian overprinting form a further subdivision.

DESCRIPTION AND ANALYSIS OF THE METAMORPHIC PATTERN

The metamorphic facies distribution (Fig. 2) displays a striking spatial correlation with previously defined crustal units and major tectonic discontinuities, which suggests that these features were fundamental crustal entities at least during Hudsonian thermotectonism. The main crustal domains (Fig. 1) are of relatively uniform metamorphic grade. Major shear zone discontinuities coincide with abrupt metamorphic gradients; generally the less abrupt metamorphic gradients are spatially related to other kinds of domain junctions. A more detailed analysis of metamorphic relations was therefore undertaken in reference to the major crustal units.

Western Craton

Textural relationships within the southwestern part of the Western Craton (Fig. 1) indicate early (Archean) metamorphic parageneses which include perthitic and antiperthitic feldspar, orthopyroxene, clinopyroxene, blue quartz, and ilmenite (Wallis, 1970b; Lewry and Sibbald, 1977) and which predate the first recognizable deformation in the Cree Lake Zone to the east. These parageneses appear to have resulted from pyroxene granulite facies metamorphism that followed the development of gneissic layering.

The granulite facies mineral parageneses are to varying degrees tectonically modified and retrogressed to assemblages composed of recrystallized quartz and feldspar, blue-green amphibole, brown-green biotite, cummingtonite, anthophyllite, sphene, and epidote. Retrogressive mafic phases typically form aggregates, but also occur as coronas about relict orthopyroxenes and clinopyroxenes. A zone of strong recrystallization in which granulite facies assemblages have almost been eliminated is present in the Virgin River Shear Zone, along the eastern edge of the Western Granulite domain (Fig. 4). This zone, which is coincident with the strongest tectonic reworking, grades westwards to a region in which both metamorphic events are clearly recognizable. Within the shear zone several periods of movement have been identified, the latest resulting in further localized retrogression (lower greenschist facies) of earlier formed Hudsonian retrograde parageneses (lower amphibolite facies) (Fig. 5). A large body of porphyritic granite was emplaced after cessation of the main movements in the shear zone, but prior to those associated with the late stage retrogression. Early Hudsonian retrograde metamorphism was apparently uniform throughout the southern part of the Western Craton and produced lower amphibolite facies assemblages, similar to those that characterize the western Virgin River domain. The retrograde metamorphism and tectonic modification was mainly contemporaneous with the first deformation farther east (Lewry and Sibbald, 1977).

A similar metamorphic history is recorded in that part of the Western Craton lying north of the Athabasca

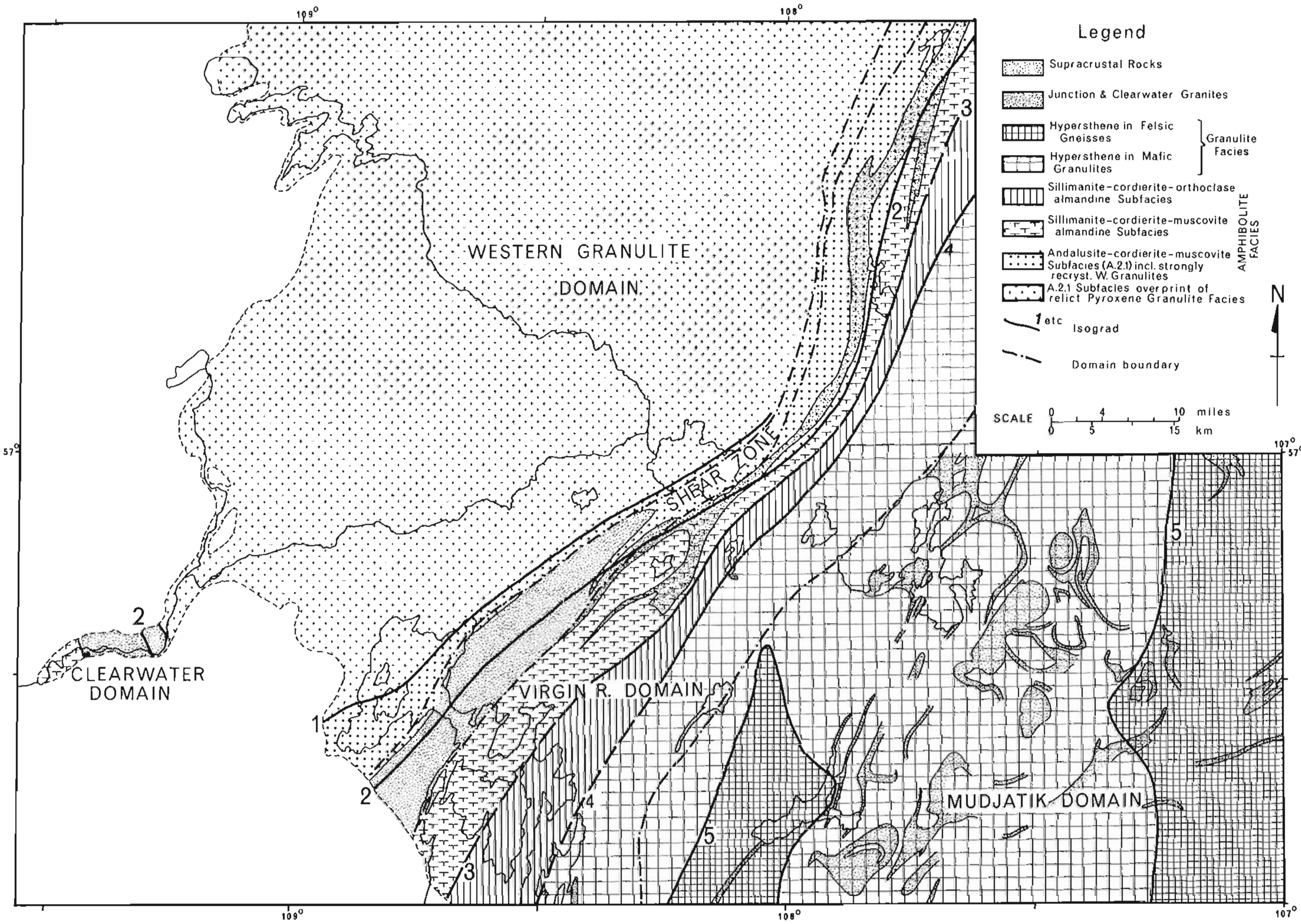

Figure 4. Metamorphic variations in the southwestern part of the Canadian Shield of Saskatchewan, showing the relationship of metamorphism to the Virgin River Shear Zone. Terminology after Winkler (1967). Figure after Lewry and Sibbald (1977). Numbering of isograds is as follows:

1. Zone of strong Hudsonian recrystallization and destruction of Archean pyroxene granulite assemblages.
2. 1st sillimanite isograd: andalusite ⇌ sillimanite.
3. 2nd sillimanite isograd muscovite + quartz ⇌ sillimanite + K feldspar + H_2O.
4. First appearance of hypersthene in mafic granulites.
5. First appearance of hypersthene in granitoid rocks.

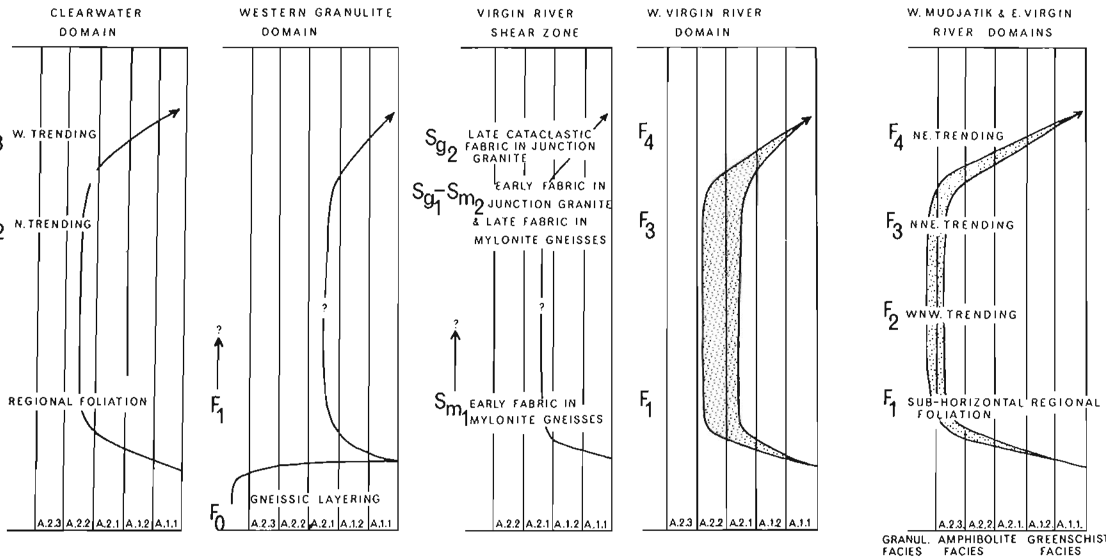

Figure 5. Temporal and spatial variations in metamorphic grade within the southwestern part of the Canadian Shield of Saskatchewan. F_1, F_2, Sm_1 etc. refer to different Hudsonian deformation episodes and F_0 to the Kenoran tectonism, recognized in the domains shown in Figure 4. Temporal relations between deformation events are illustrated by position on the vertical (time) axis. Metamorphic grade (terminology after Winkler (1967)) is depicted by the horizontal axis. The curves show the variation of metamorphic grade with time (deformation event); the striped areas show spatial variation in grade within a particular domain. Figure after Lewry and Sibbald (1977).

Formation. For example Johnston (1960, 1961) reported the existence of pyroxene granulite facies parageneses north of the Grease River Belt (Fig. 1) which are retrogressed to amphibolite facies assemblages within and to the south of this zone. The Grease River Belt, typified by both faulting and mylonite development, is identified as a linear zone of Hudsonian reworking (Beck, 1969) and can be considered analogous to the Virgin River Shear Zone, or to the Clearwater domain with which it shows apparent continuity (Wallis, 1970a; Lewry and Sibbald, 1977). Similar polymetamorphic relationships, typical of other areas in the northern part of the Western Craton (Koster and Baadsgaard, 1970; Koster, 1971; Baer, 1969), are considered to result from Hudsonian overprinting. Present data suggest that the Hudsonian overprinting was of lower grade and of a lower pressure facies series than the preceding Kenoran metamorphism. North of the Athabasca Formation kyanite occurs in little reworked Archean granulite facies areas (Johnston, 1960), whereas cordierite is restricted to zones of strong Hudsonian activity both within and to the east of the Western Craton. The cordierite-bearing granulite facies crystalline rocks forming the core of the Carswell Structure (Pagel, 1975; Herring, 1976), which are interpreted as the product of Archean metamorphism, appear anomalous in terms of the overall pattern of this metamorphism and merit further investigation.

Within the Clearwater domain (Fig. 5) a major phase of amphibolite facies recrystallization postdating the last major deformation event (F_2) is inferred from typically granoblastic fabrics. Metamorphic grade appears to have been somewhat higher than the lower amphibolite facies which characterizes the adjacent parts of the Western Craton. The eastern junction between the Clearwater domain and the Western Granulite domain (Fig. 4) was a site of granite intrusion and late tectonic movements with associated local retrogression. Similar relationships have been described in the Virgin River Shear Zone.

Cree Lake Zone

The greater part of this zone is characterized by upper amphibolite-granulite facies assemblages. Sillimanite occurs as either fibrolite sheafs, or as quartz-sillimanite knots (faserkiesel), lying with their long axes in the plane of the first foliation. Cordierite and garnet crystals sometimes appear 'flattened' in this foliation and are bent around the hinges of later folds. Foliated, folded, and boudinaged small scale granitoid segregations indicate that extensive 'in situ' anatexis may have predated and accompanied the first deformation. Field data suggest, therefore, that the earliest deformation throughout much of the Cree Lake Zone took place under P-T conditions at least as high as those corresponding to the upper amphibolite facies.

Similar, or possibly higher grade, upper amphibolite to granulite facies conditions prevailed in this area during a period of mineral growth and recrystallization which was synchronous with, and outlasted the third deformation, and resulted in the almost ubiquitous development of granoblastic fabrics. Mafic granulites are typically composed of granoblastic assemblages containing plagioclase, hornblende, diopside, and hypersthene. Hypersthene is widespread in granitic and pelitic gneiss only in the more central parts of the Cree Lake Zone.

It is probable that the high grade conditions achieved early in the deformational history were maintained at essentially the same level throughout and beyond the close of the major deformational events. Accompanying anatexis was effectively 'in situ' as there is little evidence of large scale intrusion.

There is a gentle decrease in metamorphic grade outwards across the core of the Cree Lake Zone but grade decreases rapidly within a few kilometres of the margins, marked by the Virgin River and Needle Falls shear zones. Passing into the Virgin River Shear Zone a decrease from granulite facies to lower amphibolite facies takes place in about 13 km. Andalusite, muscovite, staurolite, garnet, and cordierite are present in pelite of the Virgin Schist Group in the Virgin River Domain north of Careen Lake (Wallis, 1970b; Lewry and Sibbald, 1977). This change is illustrated in Figure 4 in terms of Winkler's (1967) subdivisions of the Abukuma facies series, although in some low grade pelite the presence of garnet (almandine?) and cordierite, or garnet alone may indicate a somewhat higher pressure facies series. Relationships between deformation and mineral growth suggest no marked changes in position of mineral isograds with time, a condition more easily reconciled with a continuous, rather than a discontinuous, metamorphic process. Temporal metamorphic variations within the different domains discussed above, are summarized in Figure 5.

The metamorphic gradient at the eastern margin of the Cree Lake Zone is less well known. Assemblages characteristic of the lowermost amphibolite facies are reported in rocks adjacent to the Needle Falls Shear Zone in a number of areas (Money, 1965; Scott, 1970, 1973; Karup-Möller, 1970) and a westward increase in grade is implied in descriptions of map units. For example, west of Pendleton Lake an increase in grade from lower to upper amphibolite facies takes place over a distance of 8 km (Scott, 1973), and at the southern end of the Needle Falls Shear Zone a similar transition from lower amphibolite to upper amphibolite-granulite facies occurs over 13 km (Munday, in prep). Somewhat higher grade middle amphibolite facies assemblages are reported immediately adjacent to the Shear Zone elsewhere (Ray, 1974, 1975; Gilboy, 1975), indicating that the structure bears a gently cross cutting relationship to the metamorphic isograds (Fig. 2).

Studies of relationships between deformation and metamorphic mineral growth (Wallis, 1971; Ray, in prep.; Gilboy, in prep.) indicate two main episodes of mineral growth accompanying the first and postdating the second major deformation phases commonly recognized in the Wollaston domain. A third phase of minor folding is generally associated with retrogression (Wallis, 1971; Scott, 1973). The conditions of metamorphism indicated by the mineral parageneses developed during these two episodes of growth appear similar.

Basement inliers along the eastern margin of the Cree Lake Zone locally display granulite facies mineral parageneses* in which hypersthene is sporadically associated with ribbon perthite, clinopyroxene, and biotite. These assemblages contrast with upper amphibolite facies assemblages in the surrounding Wollaston Group supracrustals. This relationship, particularly well shown by granitoids of the Pederson Lake complex and the surrounding supracrustals led Money (1966) to suggest that the granulite facies parageneses might date from an earlier (Archean) metamorphic event. However, the apparent higher grade of the basement inliers may also derive from lower initial water content, or as a result of the evolution of such bodies as mantled gneiss domes.

These features suggest relatively flat lying to gently domed isograds which steepen sharply at the western and eastern margins of the Cree Lake Zone. Temporal relationships between deformation and metamorphism are similar throughout and indicate that the metamorphic pattern, fixed at onset of Hudsonian deformation, persisted throughout the major part of the deformation history. This interpretation of thermal evolution is consistent with that proposed for the structural development of the zone (Sibbald, et al., 1977).

Rottenstone Complex

The metamorphic evolution of the Rottenstone Complex is not as well defined as that of the Cree Lake Zone. Key indicator minerals and mineral assemblages are restricted to metasedimentary rocks occurring as rafts and xenoliths within the western plutonic belt and to palaeosomal layers in the eastern migmatite. Two main periods of metamorphic mineral growth are recognized, the first occurred during and after the first deformation event and the second during and after the second deformation event (Ray, in prep.; Gilboy, in prep.; Lewry, in prep.). The first main period of metamorphism accompanied the early intrusive activity in the western plutonic belt and the main episode of migmatization in the migmatite belt. Sillimanite, biotite, garnet, and also, in some areas, muscovite define the first foliation and are presumed to be synchronous with development of the structure. Amphibolite facies conditions straddling the second sillimanite isograd are inferred for this period. During the second main period of mineral growth, essentially similar conditions appear to have existed although a somewhat lower pressure may be indicated by the local development of cordierite.

Emplacement of the Wathaman Batholith occurred during or after the second phase of metamorphism. Fabrics within the batholith and in the bounding Needle Falls and Parker Lake shear zones indicate deformation under relatively elevated temperatures after emplacement. Thus schistosities are marked by preferred orientations of biotite and hornblende. Late movements along the bounding shear zones, and along northerly trending fractures, associated with open folds in the Tabbernor Belt, were accompanied by local retrogression to the greenschist facies.

In the Northern Intrusive Complex (Fig. 1), diagnostic metamorphic indicators are sparsely developed. Mineral parageneses in the Campbell River Group (Lewry, 1976) just north of the Parker Lake Shear Zone and in small supracrustal remnants elsewhere suggest that the maximum grade attained was lower amphibolite facies. However Ray (in prep.) recorded an assemblage containing cordierite, garnet, and sillimanite in a small pelitic remnant immediately north of the Parker Lake Shear Zone and proposed a low pressure middle to upper amphibolite facies metamorphism for the southern part of the Northern Intrusive Complex.

Southeastern Complex

This region, the most varied in metamorphic character in the Canadian Shield in Saskatchewan, is the least understood.

The core of the western part of the La Ronge domain is characterized by fine grained rocks, in which primary structures are commonly preserved and in which there is no evidence of anatexis. These features and the observed mineral parageneses indicate metamorphism within the upper greenschist-lower amphibolite facies. Higher temperature assemblages and injection migmatites that occur in the contact metamorphic aureoles of mesosonal-epizonal plutons (Johnston, 1970), provide examples of local polymetamorphism.

An increase in metamorphic grade to the upper amphibolite-lower granulite facies characterizes the junction of the western and eastern parts of the La Ronge domain and, farther north, the junction of the La Ronge with the Kisseynew domain (Figs. 1 and 2). As indicated on the metamorphic facies and isograd maps (Figs. 2 and 3) the former transition occurs over a short distance, the latter over a wider area. The junction zones contain several different ages of granitic segregation, indicating that anatectic processes began before the earliest deformation and continued after the close of the main tectonism. Data concerning age relationships between metamorphic mineral growth and deformation are generally lacking, but in the south Reindeer Lake area straddling the junction between the

* These define local areas of granulite facies metamorphism in the predominant upper amphibolite facies terrane which form the Wollaston domain between Foster and Wollaston lakes (Fig. 2).

La Ronge and Kisseynew domains, the following tectono-metamorphic events have been recognized (Sibbald, 1977; Lewry, in prep.). In both domains the first two deformations were accompanied by middle to upper amphibolite facies metamorphism, and the third deformation, related to movements within the northerly trending Tabbernor Belt, by retrogression and cataclasis of the high grade granoblastic mineral fabrics. The authors consider deformation and metamorphism in both domains as coeval whereas Johnston (1970), has suggested the higher grade Kisseynew metamorphism to be the younger.

The metamorphic pattern in the Glennie Lake (Fig. 1) domain is marked by a general decrease in metamorphic grade to the south and an increase to the north and northwest into the eastern La Ronge domain (Fig. 2). In the northern part of the Glennie Lake domain, middle to upper amphibolite facies metamorphism and associated plutonism accompanied and

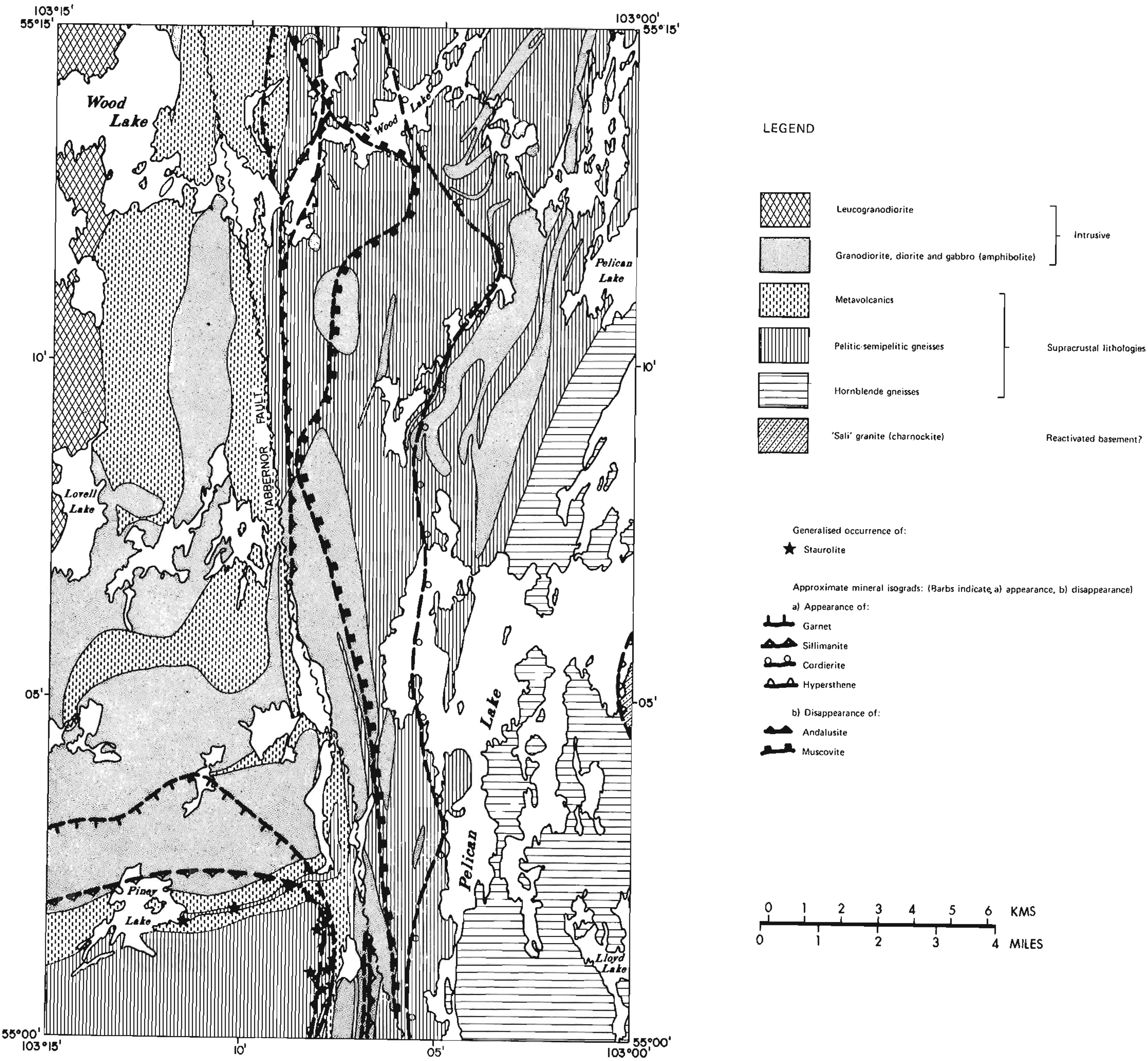

Figure 6. Metamorphic variations in the Pelican Narrows area, showing relationship to the western margin of the Tabbernor Belt. Figure after Sibbald (in press).

outlasted the first and second phases of deformation (Lewry, 1978). Locally, late growth of andalusite and muscovite postdates these events. As at the south end of Reindeer Lake the latest, third fold movements are largely postcrystalline.

Most of the southern part of the domain lies within the lower amphibolite facies, although middle amphibolite facies conditions are recorded by the appearance of sillimanite in the northeastern part of Deschambault Lake, (Figs. 2, 3). There is a general decrease in grade, to upper greenschist facies, westwards into the Wapawekka Lake region (Padgham, 1966, 1968). Polymetamorphic relationships are observed in places in the southeast of the domain. For example, early-tectonic staurolite is rimmed by late-tectonic cordierite which is coeval with andalusite porphyroblast growth (Sibbald, in press). Such relationships are in harmony with those observed by Lewry (1978) in the northern part of the domain and suggest that the latest phase of metamorphism may have taken place at a lower pressure than that in effect during the preceding metamorphic episode.

Pearson (1972), in synthesizing the structure of the Kisseynew domain, recognized three major fold generations, comprising early isoclinal folds, postdated by upright structures with northerly and northeasterly trending axial surfaces respectively. Recent studies (Pearson, 1973; Sibbald, in press; Gilboy, 1976) indicate that low pressure upper amphibolite-granulite conditions were established throughout most of the Kisseynew domain during the earliest deformation event. These high grade conditions and attendant migmatization processes are recognized to have outlasted the second folding event but their relationship to the third deformation is less certain. In some places high grade mineral fabrics are influenced by cataclasis (Cheeseman, 1959; Pyke, 1960; Pearson, 1973) suggesting that the third deformation may have outlasted the last major recrystallization. However, in other places, where similarly oriented structures occur, cataclasis was not observed and high grade metamorphism has been demonstrated to outlast deformation (Sibbald, in press). A late north-trending fourth fold event (Gilboy, 1976), possibly correlative with postcrystalline folding in the Tabbernor Belt, may represent the megascopic expression of this cataclastic episode.

There is a gentle decrease in metamorphic grade north and south out of the Kisseynew domain, and a very steep decrease to the west across the Tabbernor Belt. Detailed studies of deformation/mineral growth relationships have been carried out across one section of this belt (Sibbald, in press). The prograde sequence which typifies the Kisseynew metamorphism is marked east of the Tabbernor Fault by the replacement of andalusite by sillimanite, the disappearance of muscovite, and by the successive appearance of cordierite and hypersthene (Fig. 6). The mineral phases, while formed at different stages during the tectonic evolution of the area, are thought to comprise elements of equilibrium assemblages stable during the last period of major recrystallization following the third fold episode. This inference implies a degree of uniformity of metamorphic grade with time. An analysis of the spatial distribution of minerals of similar ages indicates that the spatial variation in metamorphism existing during the earliest deformation also persisted to outlast the main tectonic movements. The Tabbernor Belt thus appears as a significant crustal break, which was active during an early stage of Hudsonian orogenesis, as were the Needle Falls and Virgin River shear zones.

In the Flin Flon domain Stauffer and Mukherjee (1971) recognized three phases of deformation. Prograde regional metamorphism accompanied the second and strongest deformation phase. Locally developed retrogressive metamorphism is associated with late faults, which cross cut the regional metamorphic zones (Stauffer et al., 1975). Thermal metamorphic aureoles surrounding some of the larger intrusive bodies are also apparent, particularly in the lowest grade areas. A simple prograde sequence marked by the successive appearance of chlorite, biotite, garnet, andalusite, and staurolite is noted in traversing northwest into the Kisseynew gneiss. Tentative correlations have been drawn between the deformation sequence in the Flin Flon domain and that present in the Kisseynew gneiss (Pearson, 1972). Northeasterly trending third folds in the Flin Flon area are equated with similarly oriented third fold structures in the Kisseynew domain. Correlation of the two earlier deformation events in each area is less certain. If, as proposed by Pearson (1972), the main (second) deformation event in the Flin Flon domain equates with the first recognizable event in the Kisseynew domain, the metamorphic peak in the former area was coeval with the earliest period of metamorphic mineral growth in the Kisseynew gneiss. As metamorphism persisted at least locally until after the end of the third deformation in the Kisseynew gneiss (Sibbald, in press), there is some support for the view that Kisseynew metamorphism is later than that in the adjacent low grade areas (Johnston, 1970).

DISCUSSION AND SUMMARY

In most of the domains, with the possible exception of the Flin Flon domain, near peak metamorphic conditions were attained at least as early as the first major deformation event during Hudsonian orogenesis.

Early thermal activity appears to have determined deformation style; for example, in the Cree Lake Zone the level of isothermal surfaces and attendant migmatization relative to the basement-cover contact is thought to have controlled the development of either migmatite lobes or gneiss domes (Sibbald et al., 1977) in a manner analogous to that proposed by Wegmann (1935) and Haller (1971) in the Caledonian orogenic belt of East Greenland. Such structures must have been effective in promoting thermal transport to higher crustal levels. The location of major shear zone discontinuities is also dependent on the early thermal pattern. The Virgin River, Needle Falls and Tabbernor zones may be viewed as loci of ductile to brittle movements controlled in position by strong lateral thermal gradients and consequent variations in basement mobility. The shear zones and other zones of ductile to brittle reworking in the Western Craton are the sites of late tectonic granite and granite pegmatite emplacement.

Throughout most of the region the early imposed thermal pattern apparently persisted throughout the major interval of Hudsonian tectonism. A record of a thermal plateau of extended duration, accompanied or punctuated by deformational episodes, is found in most domains.

In the greater part of the Cree Lake Zone, uppermost amphibolite to granulite facies conditions persisted beyond the close of the last significant deformation episode and produced typically granoblastic, annealed fabrics. In the western part of the zone somewhat lower grade conditions outlasted the third deformation, though here tectonism continued beyond the decline from the thermal plateau to produce a fourth generation of minor fold structures and late movements in the Virgin River Shear Zone. An analogous situation is evident in the eastern part of the Wollaston domain.

Early middle amphibolite to granulite facies metamorphism extended at least beyond the close of the second deformation in the Rottenstone, eastern La Ronge, Kisseynew, and northern parts of the Glennie Lake domain. In some parts of the Kisseynew domain the northeasterly trending major folding associated with the third deformation may also predate the decline in temperature. In the northern parts of the Rottenstone, La Ronge, and Kisseynew domains a late northerly trending fold event related to movements in the Tabbernor Belt certainly postdates peak metamorphic conditions and is associated with retrogression.

The Flin Flon, western part of the La Ronge, and the southern part of the Glennie Lake domains appear to have reached and maintained upper greenschist to lower amphibolite facies conditions, possibly with a shorter interval of time at their plateau temperature. In these areas meso-epizonal plutonism produced both spatial and temporal variations in metamorphic facies.

The general persistence of the Hudsonian thermal patterns with time is supported by evidence from textural studies of deformation/mineral growth relationships in regions where steep metamorphic gradients exist, as for example adjacent to the Virgin River Shear Zone and in the Tabbernor Belt. Regions of polymetamorphism appear to be those characterized by extensive mesozonal to epizonal plutonism, in which heat was transported, at least in part, by discrete magmatic pulses.

Mineral parageneses throughout the Canadian Shield in Saskatchewan and throughout most of Hudsonian thermotectonism are consistent with low pressure (Abukuma) to low pressure-intermediate facies series metamorphism (Winkler, 1967), implying high thermal gradients and relatively shallow crustal depths of orogenic activity.

The crustal subdivisions, or lithostructural domains, forming this part of the Canadian Shield are distinguishable also as metamorphic entities. Variations in lithology, structural style, metamorphism, and associated magmatic activity may in part be explicable in terms of exposure of different crustal levels but probably they are largely due to lateral variations in early established Hudsonian thermal patterns, which in turn reflect major contrasts in crustal structure and evolution. In this context the Needle Falls Shear Zone, recognised here as a fundamental crustal feature, has been proposed as a major suture zone (Camfield and Gough, 1977). This structure is, however, the locus of low pressure (andalusite-bearing) metamorphic assemblages; the high pressure assemblages characteristic of suture zones are absent. Although higher thermal gradients and possibly thinner crust might have produced metamorphic effects that differ from those found in Phanerozoic mobile belts (Dewey and Burke, 1973) we think, on the basis of the metamorphic evidence, that if a major plate tectonic suture occurs in the Hudsonian mobile belt we must look farther east than the Needle Falls Shear Zone, and possibly outside Saskatchewan.

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THE TRANSITION FROM LOW TO HIGH GRADE METAMORPHISM IN THE KISSEYNEW SEDIMENTARY GNEISS BELT, MANITOBA

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Bailes, A.H. and McRitchie, W.D., The transition from low to high grade metamorphism in the Kisseynew sedimentary gneiss belt, Manitoba; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 155-178, 1978.

Abstract

The Kisseynew Sedimentary Gneiss Belt lies near the southeast corner of the Churchill Province, in Manitoba. It is a large, east-trending belt of Aphebian metasedimentary rocks, composed largely of greywacke, shale and derived paragneisses and migmatites.

Metamorphic conditions in the Kisseynew Sedimentary Gneiss Belt range from low grade, on its south and north margins, to high grade, in its centre. Several major isograd reactions, based on topologic discontinuities in Thompson AFM projections, are recognized from the margin to the centre of the belt:

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

chlorite + muscovite + staurolite + quartz = sillimanite + biotite + H_2O*;*

staurolite + muscovite + quartz = sillimanite + garnet + biotite + H_2O*;*

muscovite + plagioclase + quartz + H_2O *= sillimanite + melt;*

biotite + sillimanite + plagioclase + quartz + H_2O *= cordierite + garnet + melt.*

The first two reactions are responsible for the first appearance of staurolite and sillimanite, respectively, and the third controls the decomposition of staurolite in muscovite-bearing pelitic and semipelitic rocks. The last two reactions lead to partial melting and migmatization of pelitic and semipelitic rocks.

Microscopic textures in pelitic and semipelitic rocks of the Kisseynew Sedimentary Gneiss Belt, particularly paragneisses containing both sillimanite and staurolite, indicate that prograde metamorphic reactions proceeded by a mechanism of metasomatic cation-exchange.

The grade of metamorphism is highest in the centre of the Kisseynew Sedimentary Gneiss Belt and decreases rapidly along its south and north margins into the Flin Flon and Lynn Lake Greenstone belts. Partially melted and migmatitic paragneisses are widespread in the centre of the Kisseynew Belt. Pelitic gneisses are most thoroughly melted and contain anatectic segregations characterized by large and abundant xenoblasts of garnet and cordierite. Temperature during metamorphism is estimated to have exceeded 750°C and pressure is inferred to have been moderate (3.5 to 5 kb).

Résumé

La zone gneissique sédimentaire de Kisseynew est située près du coin sud-est de la province de Churchill au Manitoba. Il s'agit d'une vaste zone orientée vers l'est, de roches métasédimentaires de l'Aphébien, composées largement de grauwacke, d'argile litée, et de paragneiss et migmatites dérivés de ces matériaux.

Dans la zone gneissique sédimentaire de Kisseynew, le métamorphisme est de faible intensité sur les rebords sud et nord, et d'intensité progressivement plus élevée vers le centre de la zone. Plusieurs des principales réactions isogrades, définies en fonction des discontinuités topologiques sur les projections AFM de Thompson, ont été identifiées de la périphérie au centre de la zone:

chlorite + muscovite + grenat = staurolite + biotite + quartz + H_2O*;*

chlorite + muscovite + staurolite + quartz = sillimanite + grenat + biotite + H_2O*;*

staurolite + muscovite + quartz = sillimanite + grenat + biotite + H_2O*;*

muscovite + plagioclase + quartz + H_2O *= sillimanite + phase liquide;*

biotite + sillimanite + plagioclase + quartz + H_2O *= cordiérite + almandine + phase liquide*

Les deux premières réactions permettent d'expliquer l'apparition de la staurolite et de la sillimanite respectivement, et la troisième, la décomposition de la staurolite dans les roches pélitiques et semi-pélitiques contenant de la muscovite. Les deux dernières réactions ont provoqué la fusion partielle et la migmatisation des roches pélitiques et semi-pélitiques.

Dans les roches pélitiques et semi-pélitiques de la zone gneissique sédimentaire de Kisseynew, en particulier les paragneiss contenant à la fois de la sillimanite et de la staurolite, les textures microscopiques indiquent que des réactions métamorphiques progrades s'effectuaient suivant un mécanisme d'échange cationique par métasomatose.

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L'intensité du métamorphisme est la plus forte au centre de la zone gneissique sédimentaire de Kisseynew, et diminue rapidement le long de ses rebords sud et nord, où se trouvent les zones de roches vertes de Flin Flon et Lynn Lake. Les paragneiss ayant subi une fusion partielle, et les paragneiss migmatitiques sont fréquents au centre de la zone de Kisseynew. Les gneiss pélitiques ont subi une fusion totale, et contiennent des ségrégations anatectiques, caractérisées par la présence de nombreux xénoblastes de grande taille, constitués de grenat et cordiérite. On estime que pendant le métamorphisme, les températures ont dépassé 750°C, et que la pression a été modérée (3.5 à 5 kb).

INTRODUCTION

This paper describes the transition from low to high grade[1] metamorphism in pelitic and semipelitic gneisses of the Kisseynew Sedimentary Gneiss Belt, Manitoba. Several isograd reactions, considered typical of this transition, are described and documented in detail. Reactions, identified at high grade, involve partial melting of paragneisses.

Metamorphic phenomena observed in the File Lake area, on the south margin of the Kisseynew Sedimentary Gneiss Belt, and in the Burntwood River area, in the centre of the belt, are emphasized. The File Lake area contains a wide zone of pelitic gneisses which record the transition from low to high grade metamorphism. The Burntwood River area contains strongly recrystallized and migmatized paragneisses typical of high grade metamorphism.

Metamorphic mineral assemblages are analyzed by a method described by Thompson (1957). Many ideas presented here evolved from papers by Carmichael (1970) and Froese and Gasparrini (1975). Field data were gathered by Bailes (1970 to 1974), for the File Lake area, and by McRitchie (1970 to 1974), for the Burntwood River area, during the mapping of these areas for the Manitoba Mineral Resources Division. Additional data, for the central part of the Kisseynew Gneiss Belt, have been incorporated from a mapping program of this area carried out by the Manitoba Mineral Resources Division from 1970 to 1974 (McRitchie, Baldwin, Zwanzig and Frohlinger, in prep.). In addition, E. Froese, of the Geological Survey of Canada, made numerous constructive additions and modifications during his critical reading of our manuscript.

GEOLOGICAL SETTING

The Kisseynew Sedimentary Gneiss Belt, in the Churchill Province in north-central Manitoba (Fig. 1), is 240 km long, 140 km wide, and comprises 35 000 km^2 of highly metamorphosed sedimentary gneisses. Approximately 70 per cent of the rocks in this belt are derived from greywacke and shale; the remainder is derived mainly from arenaceous sedimentary rocks. Most of the granitic rocks in the belt are considered to be anatectic.

The Kisseynew Sedimentary Gneiss Belt is bounded on the east by Archean rocks of the Superior Province. On the west it is bounded by the Tabbernor fault zone and the granitic terrane of the Glennie Lake area. To the south and north, it is flanked by the Flin Flon and Lynn Lake Greenstone belts, respectively. Transitional lithofacies relationships between greywacke-shale derived paragneisses of the Kisseynew Sedimentary Gneiss Belt and volcanic rocks of the Flin Flon and Lynn Lake Greenstone belts, have been noted on both its south and north margins (Byers and Dahlstrom, 1954; Bailes, 1971, 1975; Zwanzig, 1976).

Rb-Sr dating studies indicate the major metamorphic, deformational and intrusive events in the Kisseynew Sedimentary Gneiss Belt, and the adjacent Flin Flon and Lynn Lake Greenstone belts, to be late Aphebian (Lowden, 1961; Lowden et al., 1963; Mukherjee et al., 1971; Anderson, 1974; Josse, 1974; Bell et al., 1975). Low initial $^{87}Sr/^{86}Sr$ ratios in metasedimentary and metavolcanic rocks of these belts imply a short crustal history and indicate that they are probably late Aphebian to middle Aphebian in age (Anderson, 1974; Josse, 1974). Pb-Pb dating studies of synvolcanic sulphide ore deposits also give late Aphebian ages (Sangster, 1972).

The Kisseynew Sedimentary Gneiss Belt is thought to represent a sedimentary basin which developed contemporaneously with volcanism within the adjacent Flin Flon and Lynn Lake Greenstone belts. At first, immature greywacke and shale, composed of slumped loose volcanic debris from the adjacent volcanic belts, were deposited by turbidity currents. This was followed by very local deposition of thin units of marlstone, rare quartzite and iron formation, succeeded, in turn, by a widespread, thick sequence of arenaceous sediments. In the Flin Flon region, the latter rocks have been interpreted by Mukherjee (1971) as a piedmont alluvial fan deposit, but their environment of deposition elsewhere in the Kisseynew Sedimentary Gneiss Belt is uncertain. Following deposition of the arenaceous sedimentary rocks, orogenesis caused up-thrusting of the entire sedimentary pile and development of a thermal high along the axis of the sedimentary basin. This metamorphic episode, the topic of this paper, formed high grade (upper almandine amphibolite to hornblende granulite facies) mineral assemblages in the centre of the Kisseynew Gneiss Belt and low grade (upper greenschist facies) assemblages in the marginal Flin Flon and Lynn Lake Greenstone belts. The main features of the metamorphic and deformational events, accompanying orogenesis, are summarized in Table 1.

REGIONAL METAMORPHIC ZONATION

A prograde sequence of critical metamorphic minerals and mineral assemblages is developed in metamorphosed greywacke and shale strata from the margin to the centre of the Kisseynew Sedimentary Gneiss Belt. The distribution of some of the more diagnostic minerals and mineral assemblages (based on an analysis of published data and on a field spot checking program by McRitchie) is shown in Figure 2. Four grades of metamorphism are recognized and outlined. They are:

Low grade: A sequence of relatively unrecrystallized metavolcanic and metasedimentary rocks characterized by biotite and chlorite.

Medium grade: A sequence of moderately recrystallized paragneisses characterized by porphyroblasts of garnet, staurolite and, near the upper boundary, sillimanite. The lower boundary is delineated by the first appearance of staurolite. Staurolite persists to near the upper boundary of medium-grade metamorphism.

High grade A (sillimanite-biotite stable): A sequence of recrystallized and migmatitic paragneisses characterized by sillimanite and garnet. The lower boundary of this zone is defined by the decomposition of muscovite in the presence of quartz and plagioclase.

High grade B (garnet-cordierite-K feldspar stable): A sequence of coarsely recrystallized and migmatized sedimentary gneisses characterized by abundant granitic mobilizate and by garnet-cordierite-K feldspar-bearing assemblages. Partial melting of aluminous pelitic rocks,

[1] The terms low, medium and high grade, to describe conditions of metamorphism, are used according to Winkler (1976).

yielding melts containing refractory garnet and cordierite, is typical. In this zone sillimanite is no longer widespread.

The transition from low to high grade conditions (or beginning of partial melting) in the Kisseynew Sedimentary Gneiss Belt is rapid, generally occurring over a distance of 10 to 20 km. Further increase in grade of metamorphism is more gradual; this probably reflects an increase in the heat required to melt sedimentary rocks in the high grade zone.

Three aspects of the regional metamorphism of the Kisseynew Sedimentary Gneiss Belt will be dealt with:

1. the transition from low grade (greenschist) to high grade (upper almandine amphibolite facies) conditions on its south margin (based largely on work by Bailes in the File Lake area);
2. the nature of high grade metamorphism (garnet-cordierite-K feldspar variety) in its centre (based on work by McRitchie, Baldwin, Zwanzig and Frohlinger);
3. the nature of anatexis and migmatization.

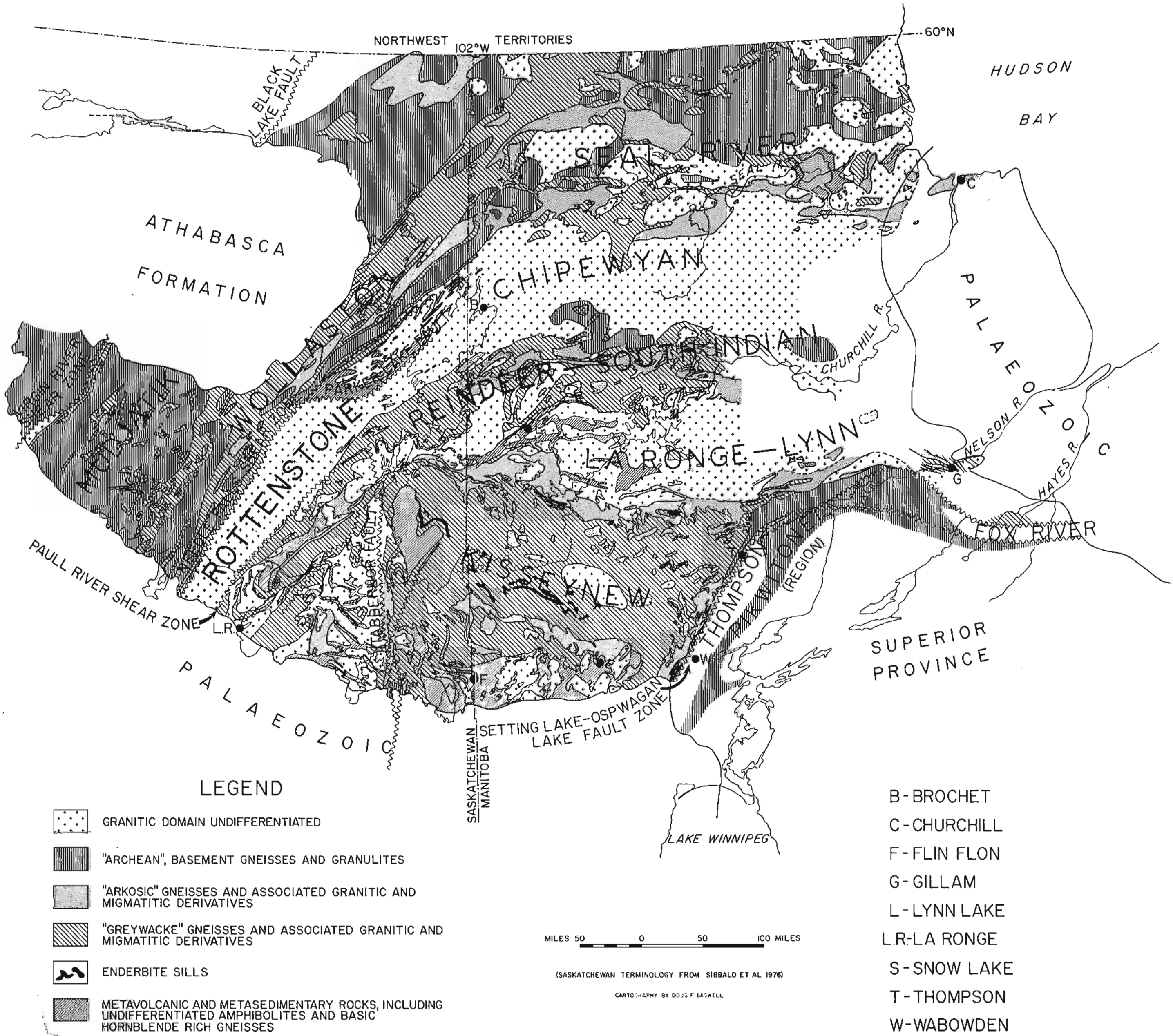


Figure 1. Major tectonic belts and lithostratigraphic units of the Churchill Province in northwest Manitoba and northeast Saskatchewan.

Table 1

Summary of metamorphic and deformational events affecting the Kisseynew Sedimentary Gneiss Belt

South Margin of Kisseynew Belt – File Lake area (Bailes, 1978, and Moore and Froese, 1972)		Central Kisseynew Belt – Nelson House-Pukatawagan area (McRitchie, Baldwin, Frohlinger and Zwanzig, in prep.)	
D_1: Large early isoclinal recumbent folds (F_1), with an axial planar schistosity (S_1).	M_2: Muscovite, chlorite and biotite define S_1. M_1 minerals are largely recrystallized and annealed during M_2.	D_1: Nappes and recumbent folds (F_1). Formation of main regional schistosity (S_1).	M_1: High grade regional metamorphism producing uppermost almandine-amphibolite to hornblende granulite facies mineral assemblages. Outlasts D_1.
D_2: Major NNE trending open shallow to steep plunging folds (F_2) which deform D_1 structures. F_2 folds locally have a prominent axial planar biotite schistosity.			
	M_2: Strong regional metamorphic event causing regional zonation of porphyroblasts. Metamorphism increases to the north from low (greenschist facies) to high (upper almandine-amphibolite facies) grade.	D_2: Synchronous (?) NNE and E-trending folds (F_2) causing basin and dome interference structures. Biotite foliation (S_2) developed locally in fold hinges.	M_2: High grade regional metamorphism (slightly lower pressure than M_1?) producing uppermost almandine-amphibolite to hornblende granulite facies assemblages.
D_3: E trending upright flexural folds (F_3), locally with axial planar biotite schistosity (S_3).			
		D_3: NNW trending folds (F_3), mainly reoriented D_2 structures. Local generation of poorly developed schistosity and zones of intense cataclastic foliation (S_3).	M_3: Hematite-sillimanite-muscovite along with overprinting S_3 cataclastic cleavages.
D_4: Major NNE trending flexure folds.		D_4: NE trending flexure folds (F_4).	
D_5: Major N and NW trending fracture systems.	M_3: Local retrogressive low grade metamorphism associated with D_5 fractures.	D_5: NNE and NW trending cataclastic faults with only minor offsets.	M_4: Low grade mineral assemblages, most prominent within D_5 structures.

THE TRANSITION FROM LOW TO HIGH GRADE CONDITIONS IN THE FILE LAKE AREA

General Statement

On the south margin of the Kisseynew Sedimentary Gneiss Belt, the metamorphic transition from relatively unrecrystallized low grade (greenschist facies) rocks to completely recrystallized and migmatized high grade (upper almandine amphibole facies) rocks is well documented in a wide zone of pelitic gneisses in the Snow Lake-File Lake area (Fig. 3). This transition was initially described by Harrison (1949). Recently, Froese and Gasparrini (1975) have proposed three isograd reactions, explaining the main mineral zonations in the Snow Lake area; and Bailes (1978) has investigated, in detail, the metamorphic rocks of the File Lake area. The results of the latter investigation are summarized here. Data from the File Lake area support the metamorphic zones and isograd reactions identified by Froese and Gasparrini (1975) and, in addition, document some metamorphic reactions for muscovite-free rocks, not discussed by Froese and Gasparrini (1975).

Prograde metamorphic phenomena in the Snow Lake and File Lake areas are interpreted, by both Bailes (in prep.) and E. Froese and J.M. Moore (pers. comm.), to be due to a single metamorphic event, M_2, which postdates an earlier, fairly insignificant, low grade metamorphic event, M_1.

Metamorphic Zones and Isograd Reactions

Introduction

Specific metamorphic reactions have been used to delineate metamorphic zones in the File Lake area, in a manner described by Carmichael (1970) and used previously, in the adjacent Snow Lake area, by Froese and Gasparrini (1975).

Four major discontinuous isograd reactions have been identified for muscovite-bearing pelitic rocks of the File Lake area:

chlorite + muscovite + garnet = staurolite + biotite + quartz + H_2O; (R.1)

chlorite + muscovite + staurolite + quartz = sillimanite + biotite + H_2O; (R.2)

staurolite + muscovite + quartz = sillimanite + garnet + biotite + H_2O; (R.3)

muscovite + plagioclase + quartz + H_2O = sillimanite + melt. (R.4)

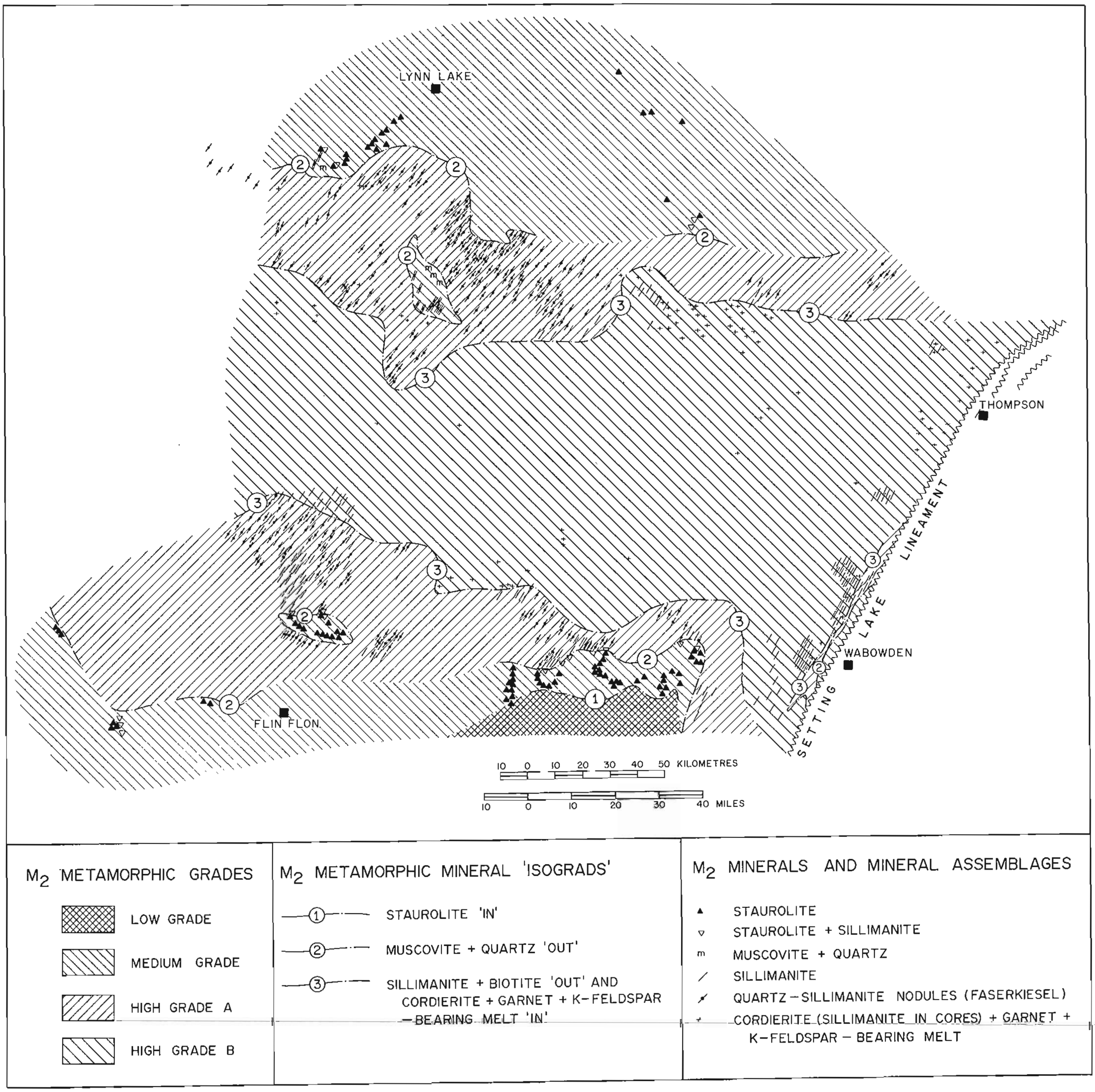


Figure 2. Regional metamorphic zonation of critical minerals and mineral assemblages in the Kisseynew Sedimentary Gneiss Belt.

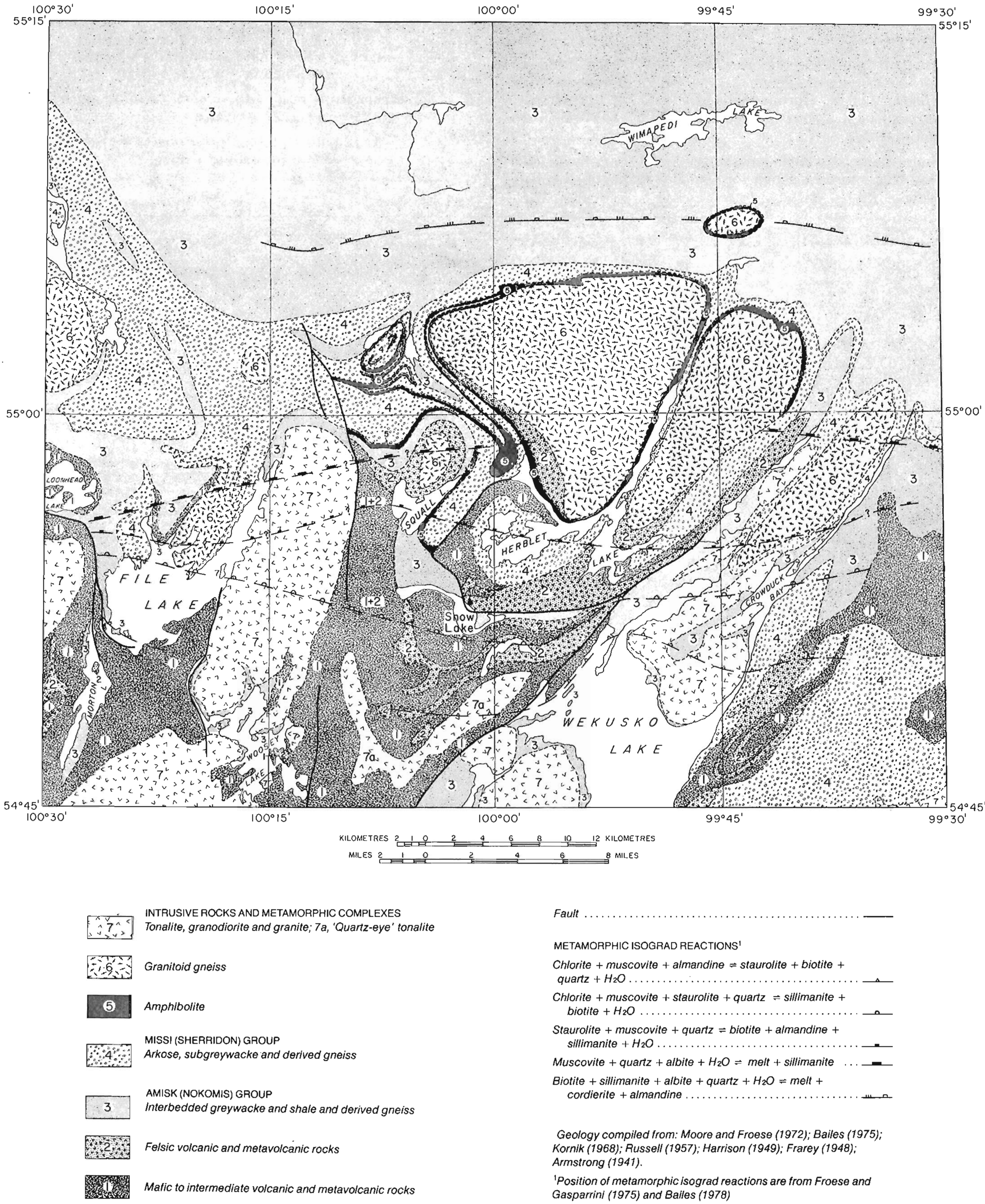


Figure 3. Simplified geological map of the File Lake – Snow Lake area (after Bailes, 1975).

These reactions are typical of the transition from low to high grade in many metamorphic belts (Guidotti, 1970 and 1974; Carmichael, 1970; Froese and Gasparrini, 1975; Pirie and Mackasey, 1978). Reactions (R.1) to (R.3) control, respectively, the first appearance of staurolite; the first appearance of sillimanite; and the decomposition of staurolite in muscovite-bearing pelitic and semipelitic rocks. Reaction (R.4) is a combination of the following two reactions:

$$\text{muscovite + plagioclase + quartz = K-Na feldspar} + Al_2SiO_5 + H_2O; \quad (R.5)$$

$$\text{K-Na feldspar + plagioclase + quartz} + H_2O = \text{melt} \quad (R.6)$$

It leads to partial melting and migmatization of pelitic and semipelitic rocks (Winkler, 1976, p. 84 & p. 309).

The four isograd reactions delineate five metamorphic zones for rocks of the File Lake area (Fig. 4):

1. chlorite-biotite zone;
2. staurolite-biotite zone;
3. sillimanite-biotite zone;
4. sillimanite-garnet-biotite zone;
5. zone of partial melting of muscovite-bearing rocks.

A description of the mineral assemblages and textures of rocks of these metamorphic zones follows.

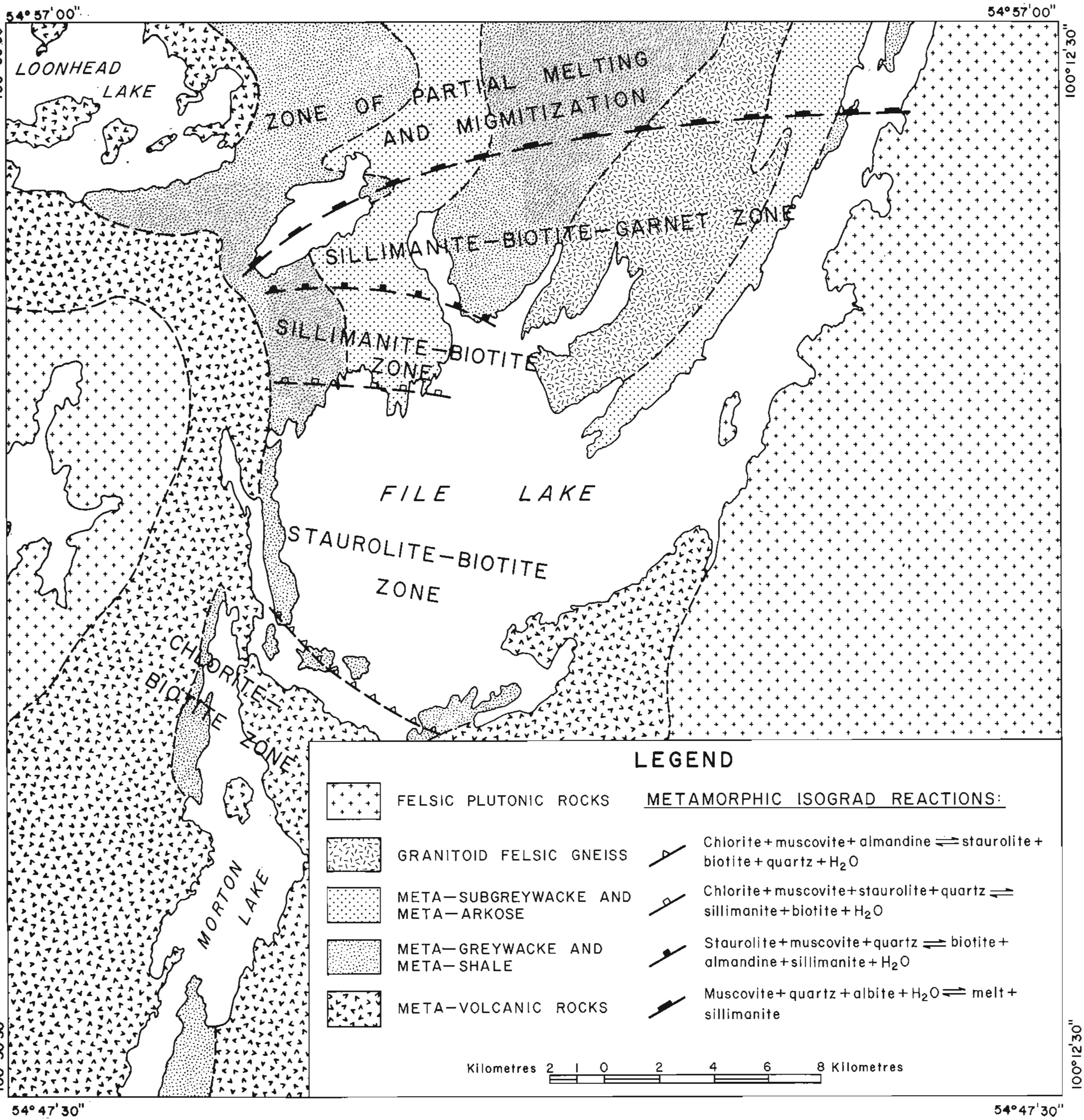


Figure 4. Disposition of metamorphic zones and isograd reactions in metagreywacke and metashale strata of the File Lake area.

Figure 5. Interbedded greywacke and shale, chlorite-biotite zone, west shore of Morton Lake. Note graded bedding in coarse sand layer and sandstone intrusion into delicately laminated shale bed. (GSC 203317-D)

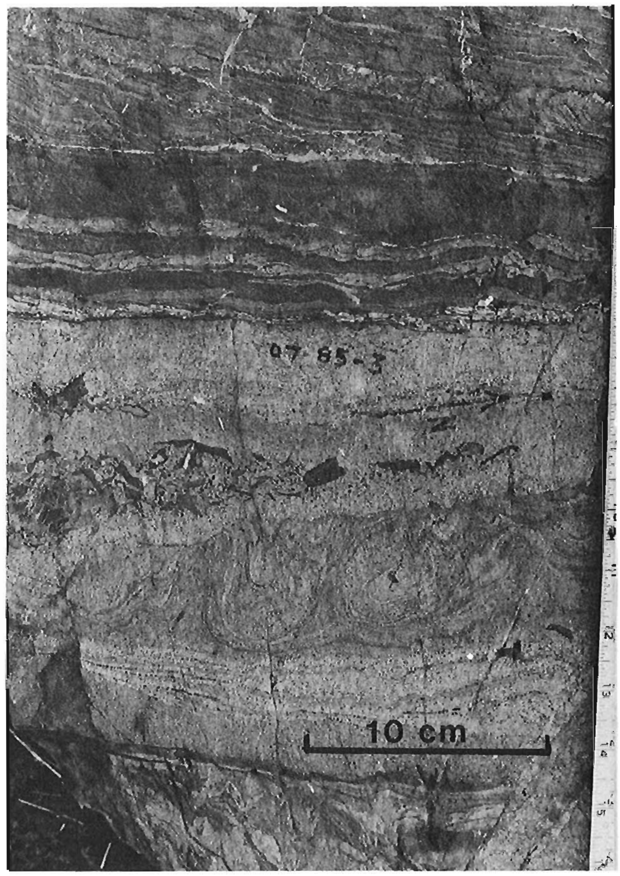


Figure 6. Shale 'rip-ups' and large convolute laminations in interbedded greywacke and shale, chlorite-biotite zone, west shore of Morton Lake. (GSC 203317-E)

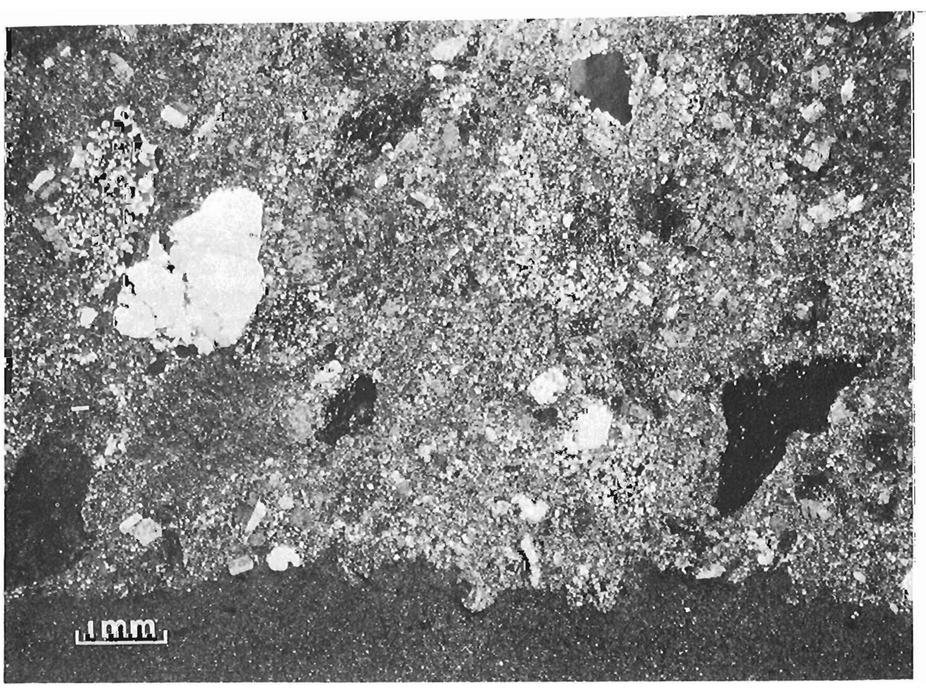


Figure 7

Photomicrograph (polarized light) of lithic greywacke, chlorite-biotite zone, east shore of Morton Lake. Note absence of strong recrystallization of clasts and groundmass matrix. (GSC 203317-F)

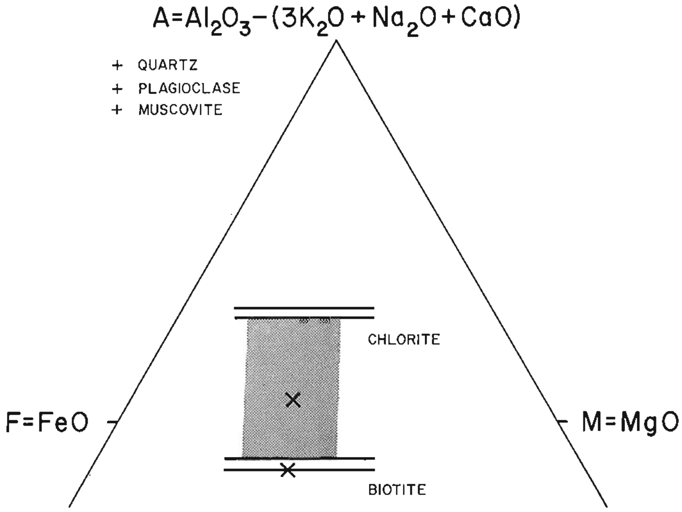


Figure 8. Schematic Thompson AFM projection of observed muscovite-bearing assemblages (shown by X) in the chlorite-biotite zone, File Lake area.

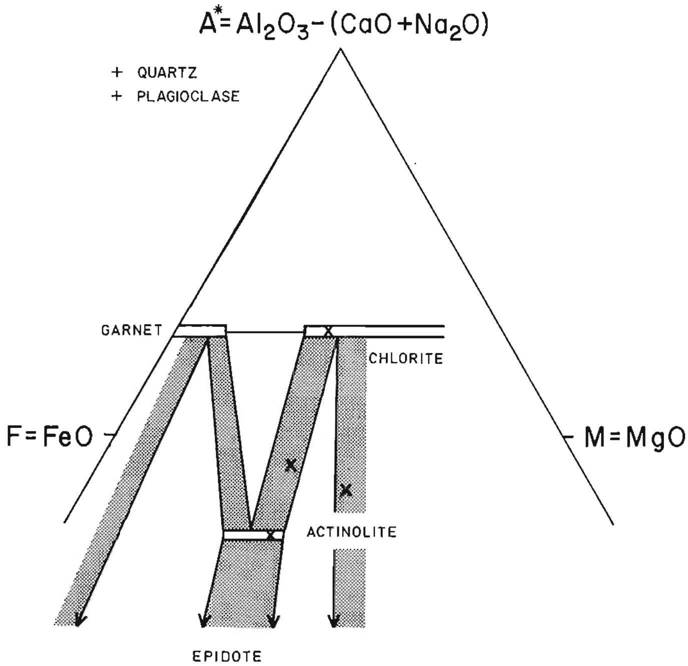


Figure 9. Schematic A*FM projection of observed muscovite-free assemblages (shown by X) in the lower chlorite-biotite zone, File Lake area.

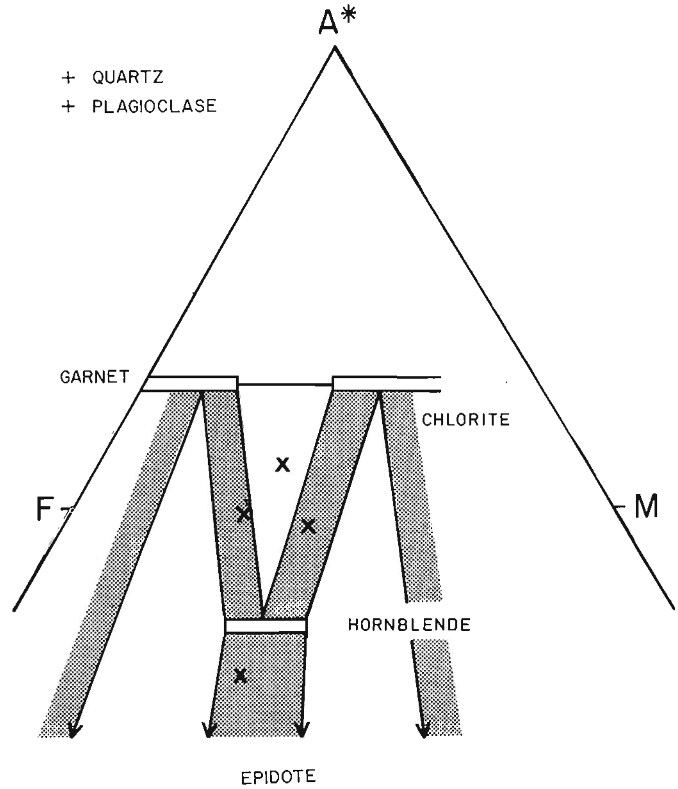


Figure 10. Schematic A*FM projection of observed muscovite-free assemblages (shown by X) in the upper chlorite-biotite zone, File Lake area.

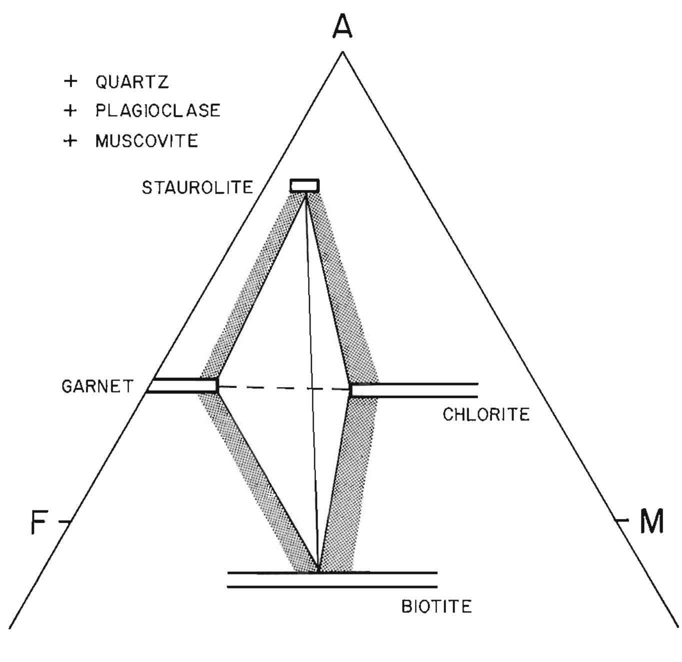


Figure 11. The discontinuous reaction at the staurolite-biotite isograd, represented on a schematic Thompson AFM projection..

Chlorite-Biotite Zone

Rocks of the chlorite-biotite zone comprise weakly metamorphosed interbedded greywacke and shale. They contain primary sedimentary structures which indicate that they were deposited by turbidity currents (Figs. 5, 6). Mineral assemblages in these rocks (Fig. 7) are characteristic of low grade (upper greenschist facies) metamorphism.

Typical mineral assemblages in K_2O-rich, muscovite-bearing rocks are shown in Figure 8, a modified Thompson AFM projection through muscovite, quartz, and plagioclase of constant composition. In this projection, $A = Al_2O_3 - (3K_2O + Na_2O + CaO)$. Typical mineral assemblages of muscovite-free rocks are shown in Figure 9, an A*FM projection through

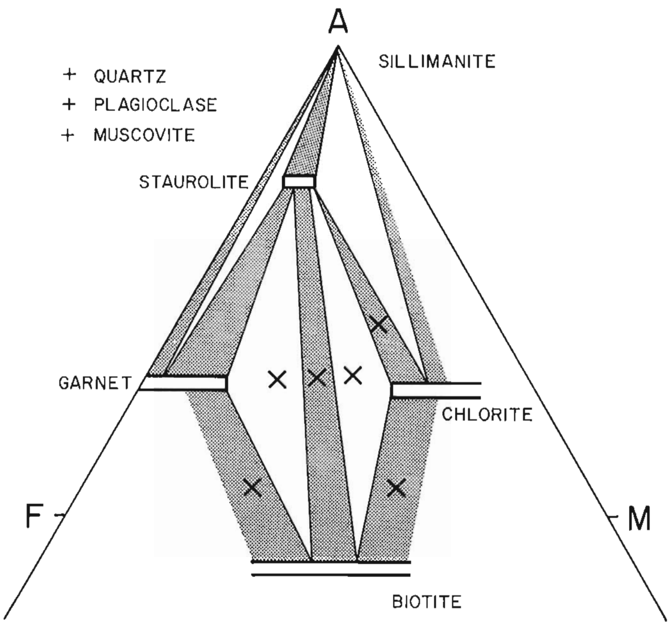


Figure 12. Schematic Thompson AFM projection of observed muscovite-bearing assemblages (shown by X) in the staurolite-biotite zone, File Lake area.

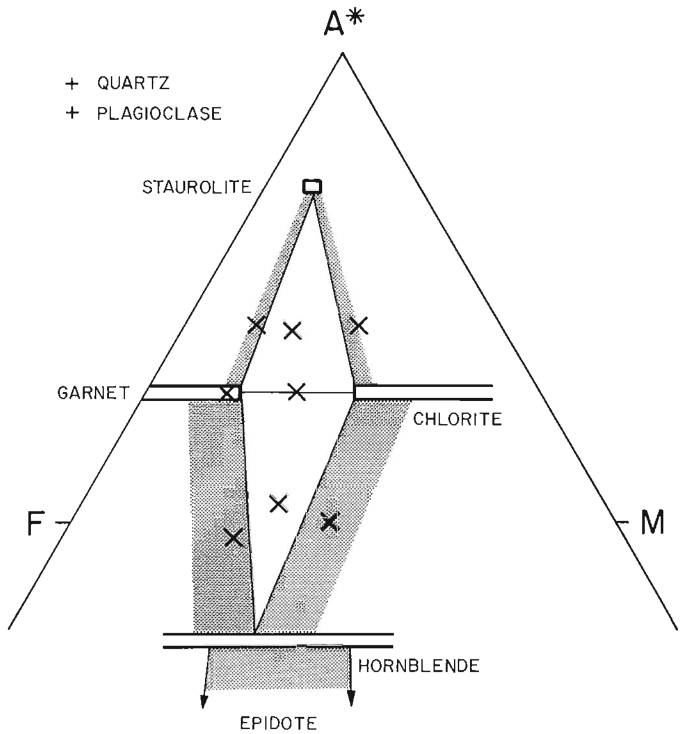


Figure 13. Schematic A*FM projection of observed muscovite-free assemblages (shown by X) in the lower staurolite-biotite zone, File Lake area.

quartz and plagioclase of constant composition, where A* = Al_2O_3 - (CaO + Na_2O). A discussion of the A*FM projection is given in a paper by Froese (1969). In K_2O-poor rocks, biotite is the only K_2O-bearing mineral and is compatible with all K_2O-free mineral assemblages shown in A*FM projections.

Near the upper boundary of the chlorite-biotite zone, garnet is found in muscovite-free rocks (Fig. 10). Its appearance is attributed to the continuous reaction

chlorite + plagioclase + quartz = hornblende
+ garnet + H_2O (R.7)

which causes the chlorite-hornblende-garnet compatibility triangle to move to more magnesian compositions.

Muscovite-bearing assemblages containing garnet were not observed in the upper part of the chlorite-biotite zone. The absence of garnet in muscovite-bearing rocks is probably due to a lack of muscovite-bearing rocks of appropriate composition.

Staurolite-Biotite Zone

The staurolite-biotite zone is separated from the chlorite-biotite zone by the discontinuous reaction

chlorite + muscovite + garnet = staurolite
+ biotite + quartz + H_2O. (R.1)

This reaction, rather than the higher grade continuous reaction

chlorite + muscovite = staurolite + biotite
+ quartz + H_2O, (R.8)

is considered to control the appearance of staurolite because garnet is present with staurolite in most rocks of the lower staurolite-biotite zone implying that garnet is also stable in muscovite-bearing rocks of the upper chlorite-biotite zone. The discontinuous reaction (R.1), referred to hereafter as the staurolite-biotite isograd, is shown in Figure 11. Assemblages which indicate that this reaction has been exceeded are: garnet-muscovite-staurolite-biotite-quartz (Fig. 12); chlorite-muscovite-staurolite-biotite-quartz (Fig. 12); and garnet-chlorite-staurolite-biotite-quartz (Fig. 13).

The transition from the lower staurolite-biotite zone to the upper staurolite-biotite zone is characterized by:

1. an increase in the abundance of staurolite, which changes from small poikiloblastic crystals to large inclusion-free idioblastic euhedral crystals (Fig. 14);
2. an increase in the range of rock compositions which contain staurolite. In the lower staurolite-biotite zone, staurolite occurs only in metashale beds but towards the upper part of the staurolite-biotite zone it also occurs widely in metasiltstone and metagreywacke beds.

These two features are probably due to the continuous reaction (R.8) which, as pointed out by Carmichael (1970), causes the staurolite-biotite-chlorite compatibility triangle to migrate towards more magnesian compositions. This reaction forms staurolite and biotite at the expense of chlorite and muscovite, causes depletion of the bank of tie lines between staurolite and chlorite, and increases the bank of tie lines between staurolite and biotite (Fig. 12). Thus the compositional field of staurolite-bearing rocks is enlarged.

In muscovite-free rocks, the following discontinuous reaction is inferred to occur about midway between the lower and upper boundaries of the staurolite-biotite zone:

chlorite + garnet + hornblende + quartz =
cummingtonite + plagioclase + H_2O. (R.9)

Mineral assemblages in muscovite-free rocks, occurring below this discontinuous reaction, are shown in Figure 13. The topology of the reaction, and the assemblages occurring above it, are shown in Figure 15.

In the Snow Lake area, E. Froese (pers. comm.) has observed one occurrence of the assemblage anthophyllite-cummingtonite-garnet-chlorite in rocks of the staurolite-biotite zone, suggesting the following equilibrium:

chlorite + garnet + cummingtonite = anthophyllite
+ quartz + H_2O (R.10)

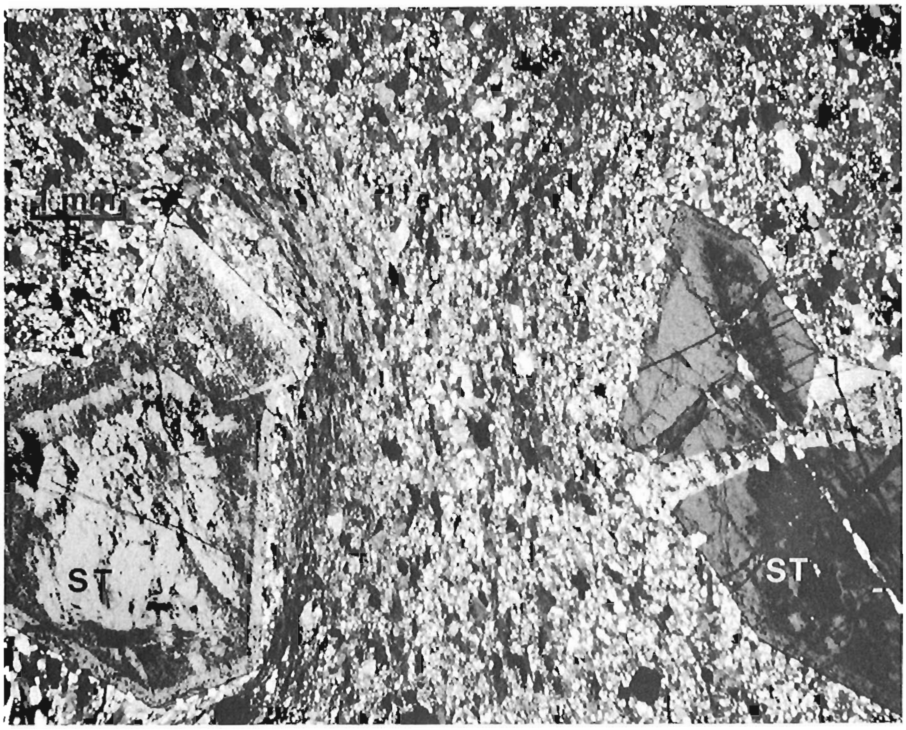


Figure 14

Photomicrograph (polarized light) of muscovite-bearing pelitic gneiss, upper staurolite-biotite zone, peninsula on west shore of File Lake. Note large euhedral staurolite porphyroblasts (ST) with growth zoning defined by inclusions. (GSC 203317-G)

Anthophyllite-bearing assemblages were not observed in the File Lake area below the sillimanite-garnet-biotite zone, but this probably reflects control by rock composition and, as in the Snow Lake area, the anthophyllite-forming reaction probably occurred in the staurolite-biotite zone.

Sillimanite-Biotite Zone

The lower boundary of the sillimanite-biotite zone is defined by the discontinuous reaction

$$\text{chlorite} + \text{muscovite} + \text{staurolite} + \text{quartz} = \text{sillimanite} + \text{biotite} + H_2O \qquad (R.2)$$

shown in Figure 16. Assemblages which indicate that this reaction has gone to completion are muscovite-sillimanite-biotite-quartz and staurolite-muscovite-sillimanite-biotite-quartz (Fig. 17). Ideally, the assemblage staurolite-chlorite-sillimanite-biotite-quartz should also occur in rocks above reaction (R.2). Its absence, in rocks of the File Lake area, is taken as an indication that its stability is restricted to a small temperature range, possibly due to the following discontinuous reaction just above the sillimanite-biotite isograd:

$$\text{chlorite} + \text{sillimanite} + \text{quartz} = \text{staurolite} + \text{cordierite} + H_2O \qquad (R.11)$$

Reaction (R.11) is inferred from an occurrence of the assemblage staurolite-cordierite-chlorite-biotite-quartz in the lower part of the sillimanite-biotite zone (Fig. 18).

The transition from the lower to upper part of the sillimanite-biotite zone is characterized by:

1. gradual depletion of staurolite, evident in a quantitative decrease and by progressive replacement of staurolite by other minerals (Figs. 23 to 26);
2. gradual increase in sillimanite, which occurs as fibrolitic knots, and a coincident loss of staurolite, chlorite and groundmass muscovite;
3. gradual increase in cordierite in very aluminous, but muscovite-free rocks.

The following three continuous reactions are interpreted to be responsible for these features:

$$\text{staurolite} + \text{muscovite} + \text{quartz} = \text{sillimanite} + \text{biotite} + H_2O; \qquad (R.12)$$

$$\text{staurolite} + \text{chlorite} + \text{quartz} = \text{cordierite} + H_2O; \qquad (R.13)$$

$$\text{staurolite} + \text{quartz} = \text{cordierite} + \text{garnet} + H_2O. \qquad (R.14)$$

As pointed out by Carmichael (1970), the first reaction causes the staurolite-sillimanite-biotite compatibility triangle to move to more iron-rich compositions and to build up the bank of tie lines between sillimanite and biotite (Fig. 17). The second reaction causes the staurolite-chlorite-cordierite compatibility triangle to move to more iron-rich compositions; it breaks down the bank of tie lines between staurolite and chlorite and leads to the discontinuous reaction

$$\text{staurolite} + \text{chlorite} + \text{quartz} = \text{cordierite} + \text{garnet} + H_2O \qquad (R.15)$$

which breaks the last staurolite-chlorite tie line (Fig. 19). The third discontinuous reaction follows reaction (R.15) and causes the bank of tie lines between staurolite and garnet and staurolite and cordierite to be depleted, ultimately leading to the discontinuous reaction

$$\text{staurolite} + \text{quartz} = \text{cordierite} + \text{garnet} + \text{sillimanite} + H_2O \qquad (R.16)$$

in the middle of the sillimanite-garnet-biotite zone (Fig. 20).

Mineral assemblages in muscovite-free rocks of the sillimanite-biotite zone in the Snow Lake area indicate somewhat higher pressure reactions than those deduced for the File Lake area (E. Froese, pers. comm.).

Sillimanite-Garnet-Biotite Zone

The lower boundary of the sillimanite-garnet-biotite zone is defined by the discontinuous reaction R.3. This reaction, referred to as the sillimanite-garnet-biotite isograd, is shown schematically in a Thompson AFM projection in Figure 24. Assemblages which indicate that this reaction has

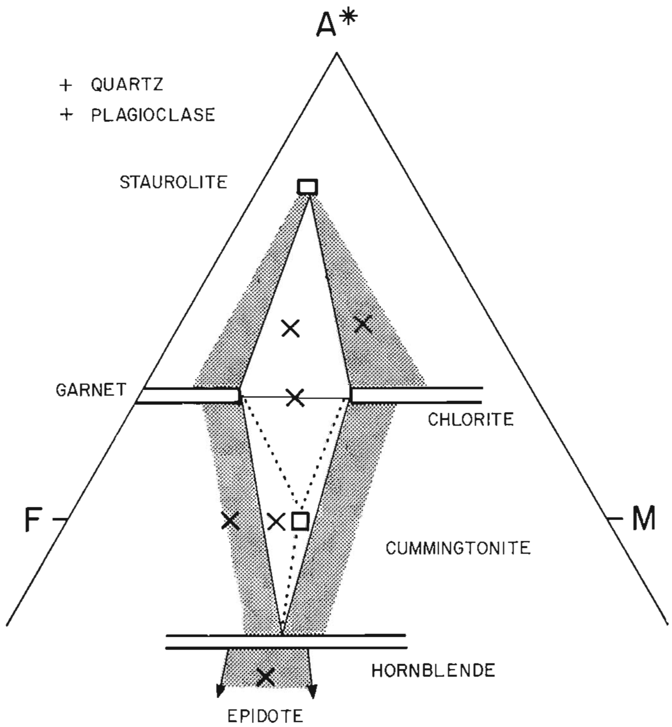


Figure 15. Schematic A*FM projection depicting the discontinuous reaction responsible for the first appearance of cummingtonite in muscovite-free rocks of the staurolite-biotite zone, File Lake area. Assemblages stable above this reaction are shown by X.

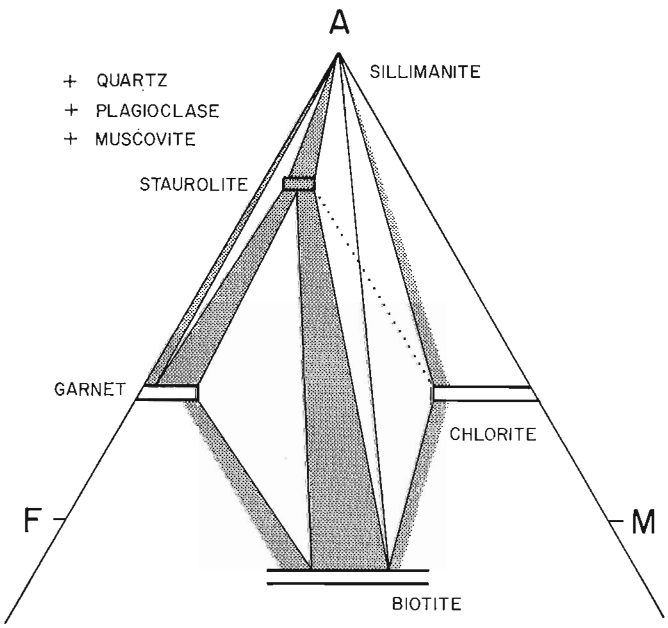


Figure 16. The discontinuous reaction at the sillimanite-biotite isograd, represented on a schematic Thompson AFM projection.

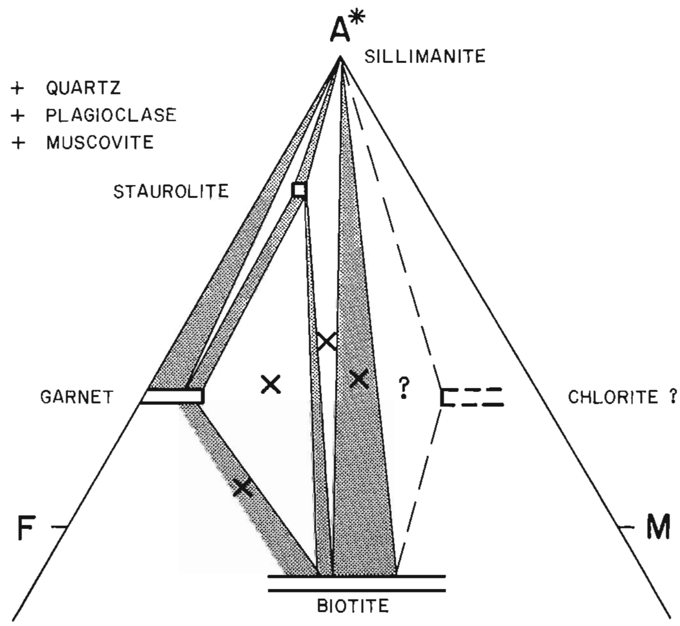


Figure 17. Schematic Thompson AFM projection of observed muscovite-bearing assemblages (shown by X) in the sillimanite-biotite zone, File Lake area.

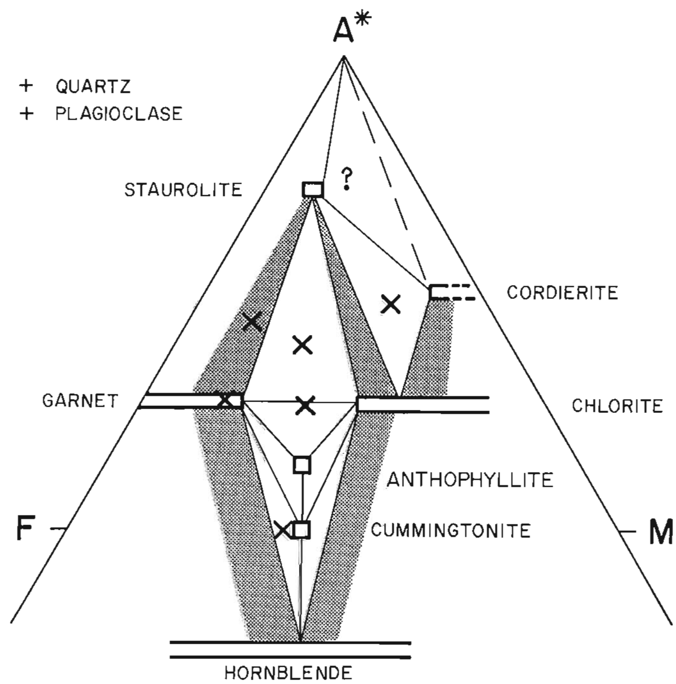


Figure 18. Schematic A*FM projection of observed muscovite-free assemblages (shown by X) in the lower sillimanite-biotite zone, File Lake area.

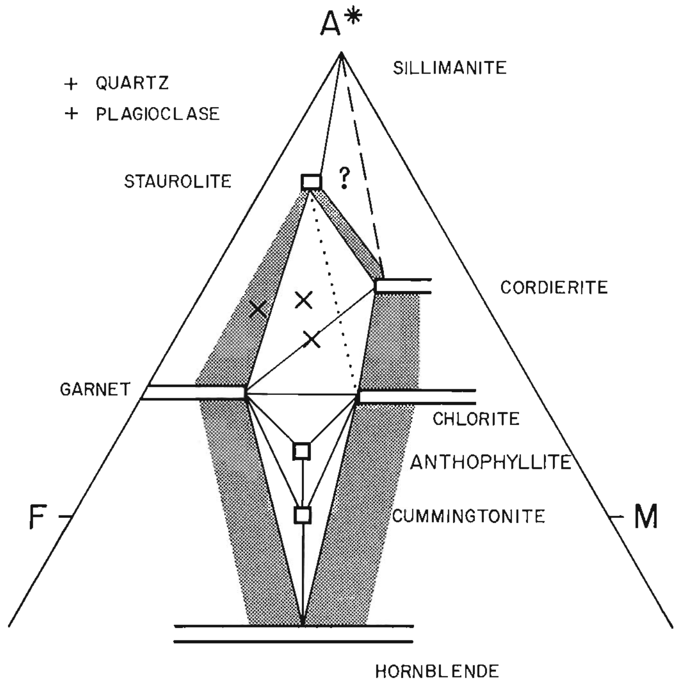


Figure 19. Schematic A*FM projection depicting the discontinuous reaction responsible for the first appearance of cordierite + garnet in muscovite-free rocks of the lower sillimanite-biotite zone, File Lake area. Assemblages stable above this reaction are shown by X.

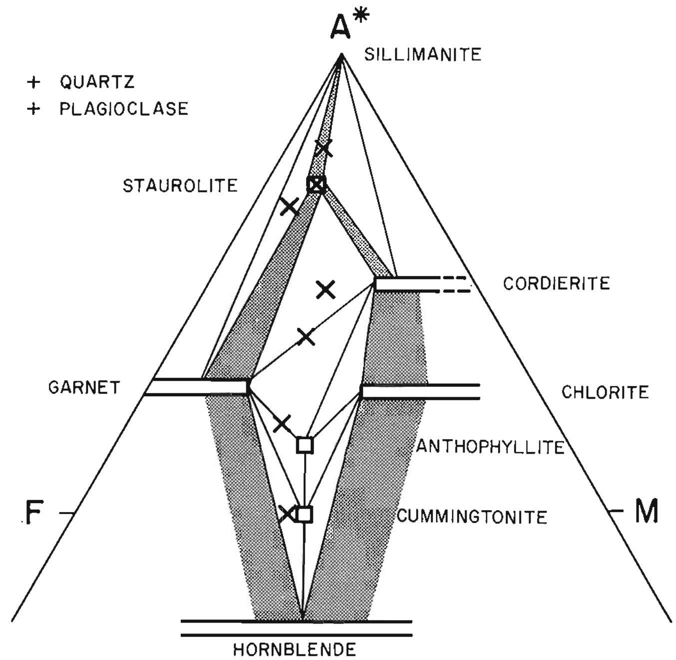


Figure 21. Schematic A*FM projection of observed muscovite-free assemblages (shown by X) in the lower sillimanite-garnet-biotite zone, File Lake area.

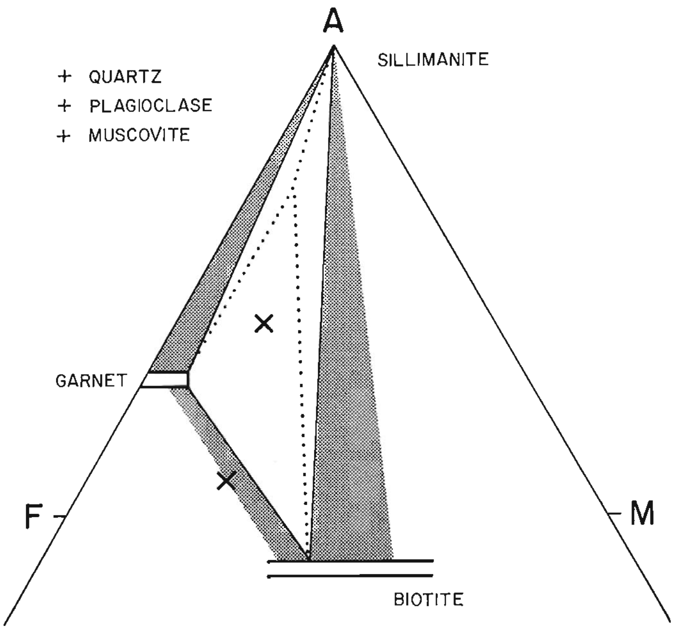


Figure 20. The discontinuous reaction at the sillimanite-garnet-biotite isograd, represented on a schematic Thompson AFM projection. Assemblages stable above the isograd are shown by X.

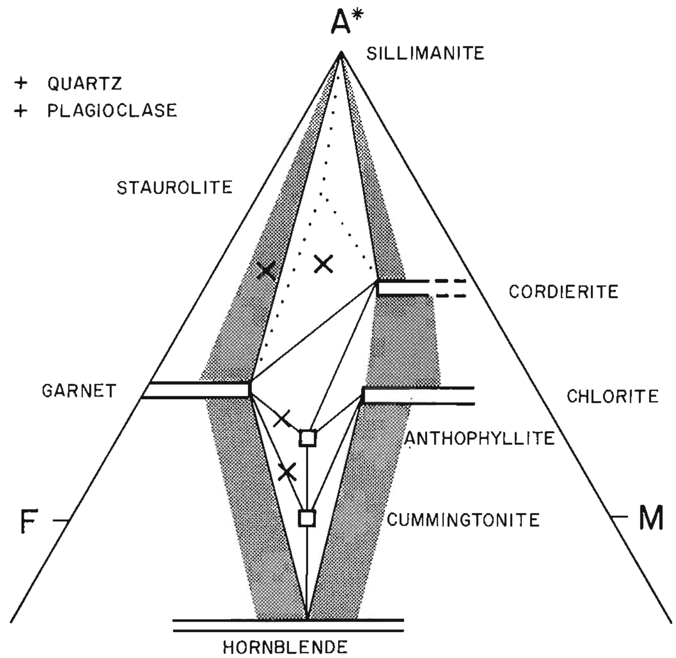


Figure 22. Schematic A*FM projection depicting the discontinuous reaction responsible for the decomposition of staurolite in muscovite-free rocks. Assemblages stable above this reaction are shown by X.

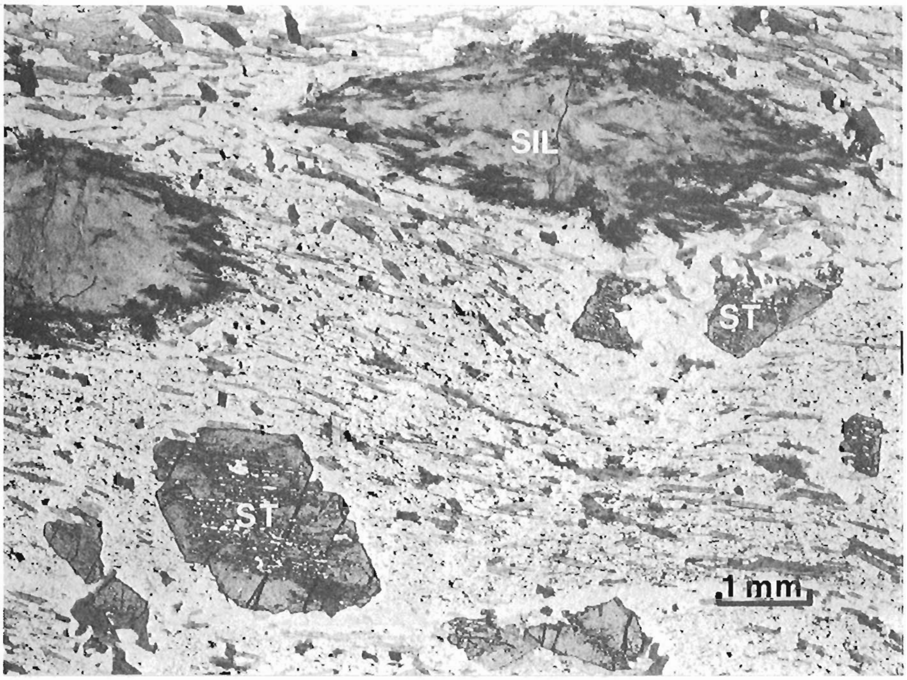


Figure 23

Photomicrograph (plain light) of muscovite-bearing pelitic gneiss, lower sillimanite-biotite zone, northwest shore of File Lake. Note large nodular aggregates of fibrolitic sillimanite (SIL) and partially corroded porphyroblasts of staurolite (ST). The sillimanite nodules are surrounded by a rim of disseminated magnetite (dark material). (GSC 203317-H)

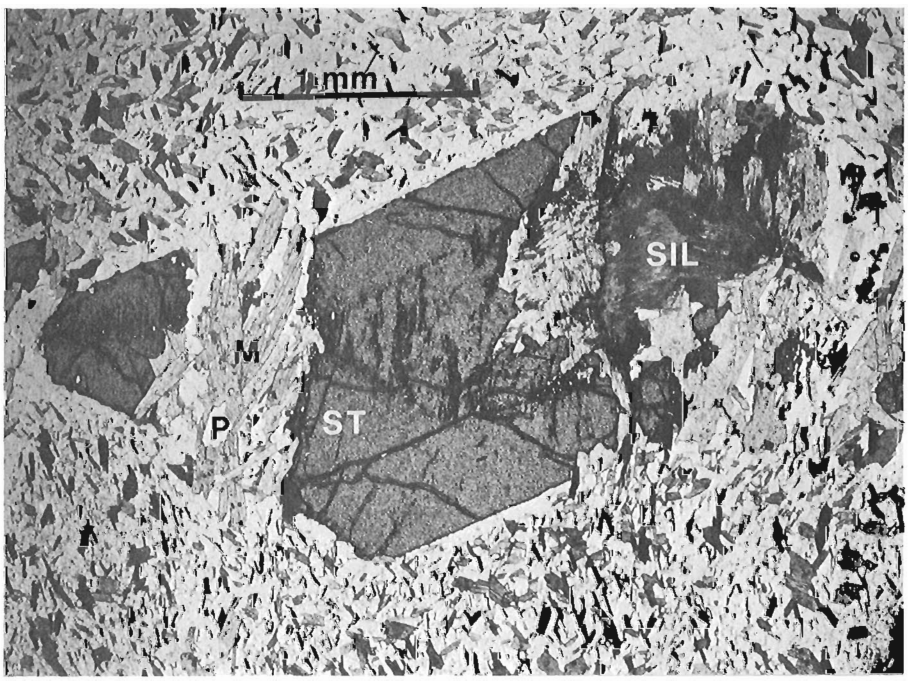


Figure 24

Photomicrograph (plain light) of a corroded porphyroblast of staurolite (ST), partially replaced by a mixture of prograde muscovite (M), fibrolitic sillimanite (SIL) and plagioclase (P), in a muscovite-bearing pelitic gneiss, lower sillimanite-biotite zone. (GSC 203317-I)

gone to completion are muscovite-sillimanite-biotite-garnet-quartz (Fig. 20) and staurolite-sillimanite-biotite-garnet-quartz (Fig. 21).

In the File Lake area, staurolite persists in muscovite-free rocks approximately 0.5 km beyond the sillimanite-garnet-biotite isograd. It is ultimately decomposed by the discontinuous reaction

$$\text{staurolite} + \text{quartz} = \text{cordierite} + \text{garnet} + \text{sillimanite} + H_2O \qquad \text{(R.16)}$$

which is shown in Figure 26. This reaction permits sillimanite to form in muscovite-free pelitic and semipelitic rocks by breaking the tie lines from staurolite to cordierite and to garnet.

Anthophyllite-bearing rocks in the sillimanite-garnet-biotite zone indicate that the discontinuous reaction (R.10) has taken place (Figs. 21, 22).

Zone of Partial Melting of Muscovite-bearing Rocks

Rocks of the migmatite zone of the File Lake area are characterized by: prominent veining by a white tonalite and granodiorite; disappearance of muscovite; coarse recrystallization (a grain size of 0.1 to 1.0 mm is characteristic compared to a grain size of 0.01 to 0.02 for the groundmass of rocks of the chlorite-biotite zone); and a decrease in the amount of sillimanite.

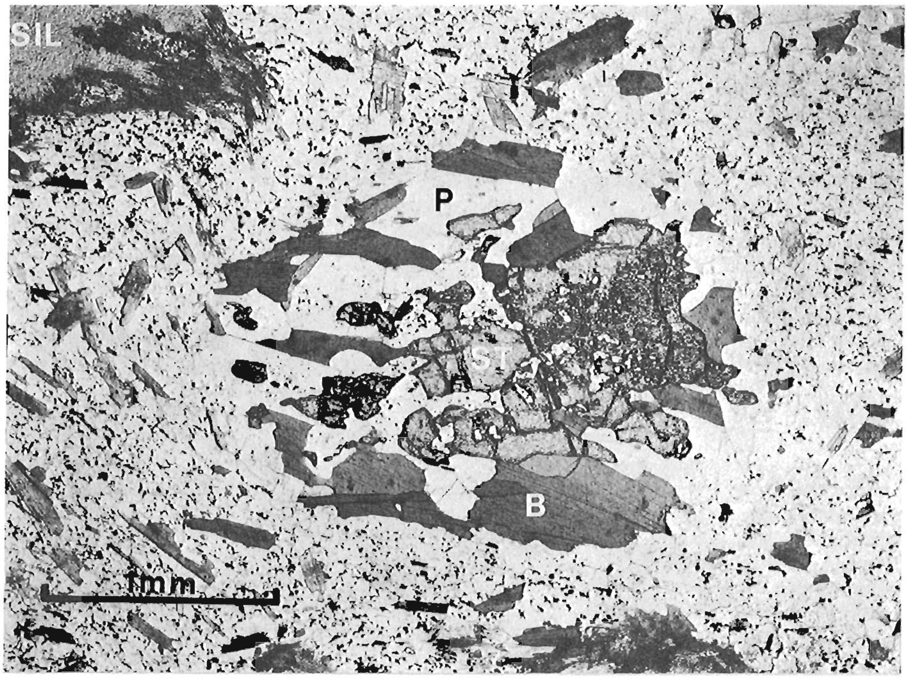


Figure 25

Photomicrograph (plain light) of a corroded porphyroblast of staurolite (ST) partially replaced by a mixture of plagioclase (P) and biotite (B), in a muscovite-bearing pelitic gneiss, middle sillimanite-biotite zone. (GSC 203317-J)

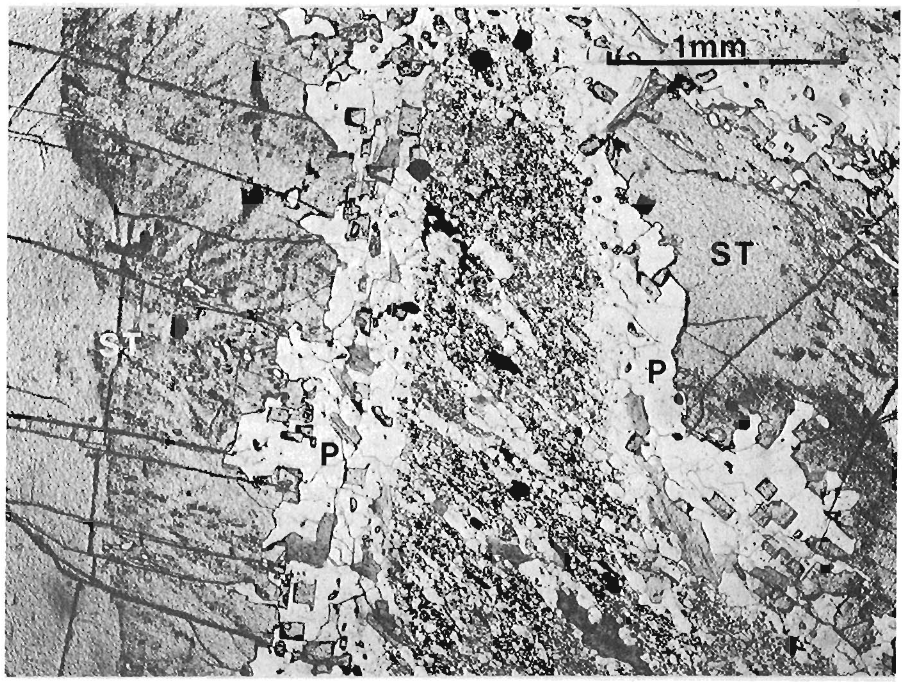


Figure 26

Photomicrograph (plain light) of staurolite porphyroblasts (ST), partially replaced by plagioclase (P), in a muscovite-free pelitic gneiss, middle sillimanite-biotite zone. (GSC203317-K)

The disappearance of muscovite and the coincident appearance of small bodies and lits of anatectic material, suggest that the lower boundary of the migmatite zone is defined by the reaction R.4. According to Winkler (1976) this reaction only occurs in rocks under conditions of moderate to high pressure.

The depletion of sillimanite observed in rocks of the migmatitic zone is, however, at odds with isograd reaction R.4 which suggests that sillimanite should be produced rather than consumed.

Textures and the Mechanism of Prograde Reactions

Introduction

One of the problems with isograd reactions based on discontinuities in topology of Thompson AFM projections is that microscopic textures often do not agree with the topologically deduced reactions (Turner and Verhoogen, 1960, p. 450; Chinner, 1961; Carmichael, 1969). This problem has also been encountered in the File Lake area, particularly in rocks of the sillimanite-biotite zone. The problem textures are:

1. replacement of staurolite by prograde muscovite (Fig. 24);
2. replacement of staurolite by prograde plagioclase (Figs. 25, 26);
3. the occurrence of sillimanite in fibrolitic knots isolated from other aluminous minerals (Fig. 23).

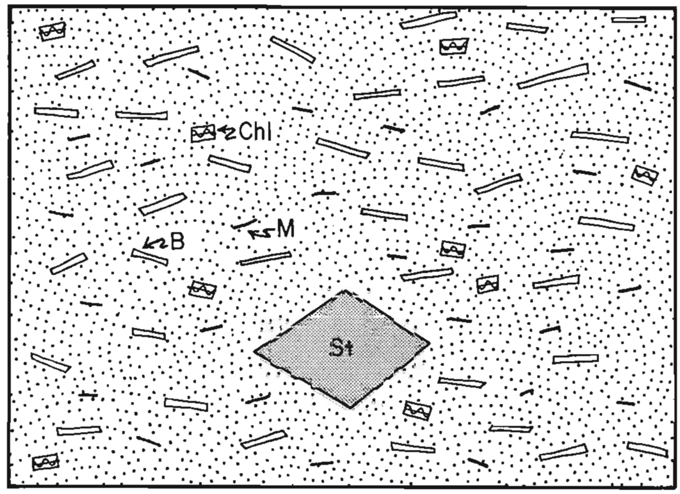
It will be demonstrated, in this section, that although these textures are consistent with the topologically determined reactions discussed previously, their reaction mechanism is more complicated than a simple transformation of reactants to products. The textures appear to be the result of diffusion-controlled cation exchange reactions (Carmichael, 1969).

Description of Textures

Plates of prograde muscovite rimming and replacing staurolite first occur in the File Lake area just above the sillimanite-biotite isograd and are found throughout the sillimanite-biotite zone until staurolite is finally consumed at the sillimanite-garnet-biotite isograd. The replacement of staurolite by muscovite is generally accompanied by the formation of large plates of biotite, coarsely crystalline plagioclase, and, in one sample, sillimanite (Fig. 24). Despite the formation of new prograde muscovite replacing staurolite, the muscovite decreases in quantity upgrade in the sillimanite-biotite zone, largely by consumption of matrix muscovite. The interpretation of the muscovite as a retrograde mineral overprinting staurolite is ruled out because its first appearance coincides with the sillimanite-biotite isograd and no overprinting of staurolite by muscovite was observed below this isograd. This phenomenon is not unique to the File Lake area, as the following description of pelitic paragneiss of the Rangeley-Oquossac area of Maine, from Guidotti (1968), demonstrates:

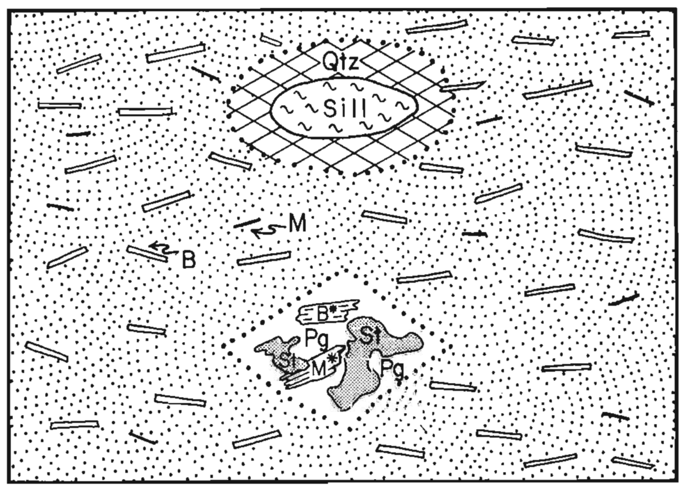
> "Coinciding approximately with the first appearance of sillimanite, staurolite in many specimens becomes anhedral with coarse laths of muscovite occurring around the outer rim. At progressively higher grades, in the lower sillimanite zone, the muscovite becomes more pronounced and the rimmed staurolite shrinks..."

New coarsely crystalline prograde aggregates of plagioclase rimming and replacing staurolite first occur just above, and appear to be related to, the sillimanite-biotite isograd. The prograde replacement of staurolite by plagioclase occurs in both muscovite-bearing and muscovite-free rocks and continues throughout the sillimanite-biotite zone (and into the sillimanite-garnet-biotite zone in muscovite-free rocks) until the staurolite is consumed. It begins as a minor corrosion of the periphery of the staurolite grains (Fig. 26) and ends as complete replacement of them. In muscovite-bearing rocks, the plagioclase is generally accompanied by both muscovite and biotite. In muscovite-free rocks, the



chlorite + muscovite + staurolite + quartz =

sillimanite + biotite + H_2O



M*	= muscovite in groundmass	St	= staurolite
M	= new muscovite plates	Sill	= sillimanite
B*	= biotite in groundmass	Dotted area	= groundmass mixture of quartz + plagioclase
B	= new biotite plates		
Chl	= chlorite		
Pg	= plagioclase		
		Qtz	= quartz

Figure 27. Typical texture and mineralogy of a muscovite-bearing pelitic gneiss below (a) and above (b) the sillimanite-biotite isograd, File Lake area.

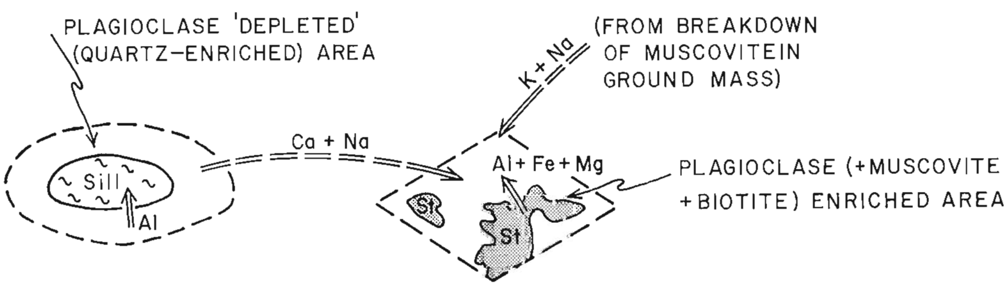


Figure 28

Main cation exchanges forming plagioclase depleted areas around sillimanite knots and plagioclase-enriched areas around, and replacing, staurolite in a muscovite-bearing pelitic gneiss of the sillimanite-biotite zone (see Fig. 27).

plagioclase occurs alone (Fig. 26) or with biotite replacing and rimming staurolite. Plagioclase replacement of staurolite is most pronounced in muscovite-free rocks, particularly in those containing cordierite.

Metamorphic nodules of sillimanite are characteristic of muscovite- and staurolite-bearing pelitic gneisses of the sillimanite-biotite zone. They comprise aggregates of fibrolitic sillimanite enveloped by quartz-rich domains and are usually completely isolated from other aluminous materials.

Interpretation of Textures

Carmichael (1969) suggested that prograde reactions in metamorphosed pelitic rocks proceed by local cation exchange reactions. He assumed that aluminum remained immobile and demonstrated that a set of cation exchange reactions, when summed over a thin section, can yield a net reaction which is equivalent to a change in topology of an AFM diagram. The textures in pelitic rocks of the sillimanite-biotite zone of the File Lake area support the cation exchange mechanism suggested by Carmichael (1969) for prograde reactions. However, they indicate that aluminum diffuses over short distances and thus is not entirely immobile as suggested by Carmichael (1969).

Typical textures and mineral assemblages of a muscovite-bearing pelitic gneiss from below and above the sillimanite-biotite isograd are compared in Figure 27. Below the isograd, staurolite occurs as well formed euhedral porphyroblasts (Fig. 14). Above the sillimanite-biotite isograd, the staurolite porphyroblasts are corroded and replaced by a mixture of plagioclase, biotite and muscovite; and aggregates of sillimanite, enveloped by a quartz-rich zone, are nucleated where previously had been a groundmass mixture of plagioclase, quartz, biotite, muscovite and chlorite (Fig. 23). Less obvious, but also important, is the absence of chlorite and the reduced amount of muscovite in pelitic gneisses above the sillimanite-biotite isograd. The main cation exchanges necessary to produce the textures in the sillimanite and staurolite domains of pelitic gneisses above the sillimanite-biotite isograd reaction are schematically shown in Figure 28.

An important feature of the cation exchange mechanism, as pointed out by Carmichael (1969), is that reactant and product phases of a topologic reaction can interact without being in physical contact. This is convincingly demonstrated for the sillimanite-biotite isograd reaction (R.2) by the textures and the interpreted cation exchange reactions depicted in Figures 27 and 28. The cation exchange reaction mechanism allows staurolite and sillimanite to react without being in physical contact and without the need for transporting relatively immobile aluminum between the two. This is largely accomplished by the dissolution of matrix plagioclase. The aluminum released from the dissolving plagioclase forms sillimanite and the unneeded Na and Ca cations combine with aluminum released from dissolving staurolite to form new coarsely crystalline plagioclase. In a similar manner, other cation exchanges take place by diffusion in response to chemical potential gradients.

The dissolution of staurolite in muscovite-free pelitic rocks of the File Lake area was accomplished by a cation exchange mechanism similar to the one which operated in muscovite-bearing rocks. Staurolite was replaced by plagioclase (Fig. 26), but cordierite, rather than sillimanite, nucleated in the groundmass. The cordierite replaced groundmass plagioclase, a process using Fe and Mg from the dissolving staurolite and releasing Na and Ca cations. Simultaneously, the staurolite was replaced by plagioclase, which used aluminum released from the dissolving staurolite and Na and Ca cations released during the replacement of groundmass plagioclase by cordierite.

HIGH GRADE METAMORPHISM IN THE CENTRE OF THE KISSEYNEW SEDIMENTARY GNEISS BELT

General Statement

The highest grade rocks of the File Lake area are characterized by partial melting and migmatization of pelitic gneisses accompanying the decomposition of muscovite. To the north, and towards the centre of the Kisseynew Sedimentary Gneiss Belt, the metamorphic grade increases gradually, and pelitic gneisses become progressively more migmatitic and enriched in granitoid mobilizate. A marked increase in degree of migmatization and amount of granitoid mobilizate occurs above the isograd that represents the reaction

$$\text{biotite + sillimanite + plagioclase + quartz} + H_2O \rightleftharpoons \text{cordierite + garnet + melt}$$

which separates the high grade A and high grade B zones shown in Figure 2.

Four separate metamorphic pulses, each with its own mineralogical and textural characteristics, have been identified in the centre of the Kisseynew Sedimentary Gneiss Belt (McRitchie, Baldwin, Zwanzig and Frohlinger, in prep.). The initial two periods of metamorphism (M_1 and M_2) are prograde metamorphic events and represent the thermal climax reached in the Kisseynew Sedimentary Gneiss Belt (Table 1). These two metamorphic events will be considered in this section. Textures produced by both metamorphic events, particularly in pelitic gneisses, indicate that cation mobility was very high. Both the M_1 and M_2 events appear to have been initiated at approximately the same temperatures, with the pressures during M_1 possibly slightly higher than those during M_2.

In general, pelitic and semipelitic gneisses have responded much more dramatically to high grade metamorphism than have psammitic gneisses, which have maintained a monotonous mineralogy and simple textures throughout high grade metamorphism (Figs. 29, 30). For this reason, the metamorphism of the pelitic and semipelitic gneisses will be emphasized.

M_1 Metamorphism

M_1 mineral assemblages indicate that high grade (upper almandine amphibolite to hornblende granulite facies) conditions were reached in the centre of the Kisseynew Sedimentary Gneiss Belt. They usually comprise relict porphyroblasts, partially converted to M_2 assemblages.

In psammites, assemblages interpreted as typical of M_1 are: garnet-cordierite-orthoclase-biotite-quartz-plagioclase; cordierite-biotite-orthoclase-quartz $\pm$ plagioclase; garnet-cordierite-biotite-quartz-plagioclase; and garnet-biotite-quartz-plagioclase. Sillimanite is observed in all these assemblages but it is not abundant and typically occurs as fine acicular needles within other minerals (with the exception of orthoclase). Accessory phases include graphite, apatite and zircon. Hypersthene has been observed in two localities in the assemblage garnet-hypersthene-orthoclase-quartz-plagioclase. Hypersthene and garnet in this assemblage occur within thin, irregularly layered, coarse grained 'granitic' segregations in a pelitic host. M_1 hypersthene also has been recorded in the assemblage hypersthene-quartz-plagioclase $\pm$ biotite in several localities within the centre of the Kisseynew Sedimentary Gneiss Belt.

In pelites and semipelites, M_1 mineral assemblages are not well preserved due to strong recrystallization during M_2. However, the assemblage garnet-cordierite-sillimanite-biotite-hercynite-plagioclase-quartz $\pm$ potassium feldspar is considered typical of the pelites and semipelites. Accessory minerals in the pelites and semipelites include apatite, graphite, zircon and magnetite. The M_1 garnet and cordierite

Figure 29

Highly metamorphosed interbedded pelitic and psammitic gneiss, high grade part of the Kisseynew Sedimentary Gneiss Belt. Note preferential partial melting of pelitic gneisses and concentration in them of large garnet porphyroblasts. (GSC 203317)

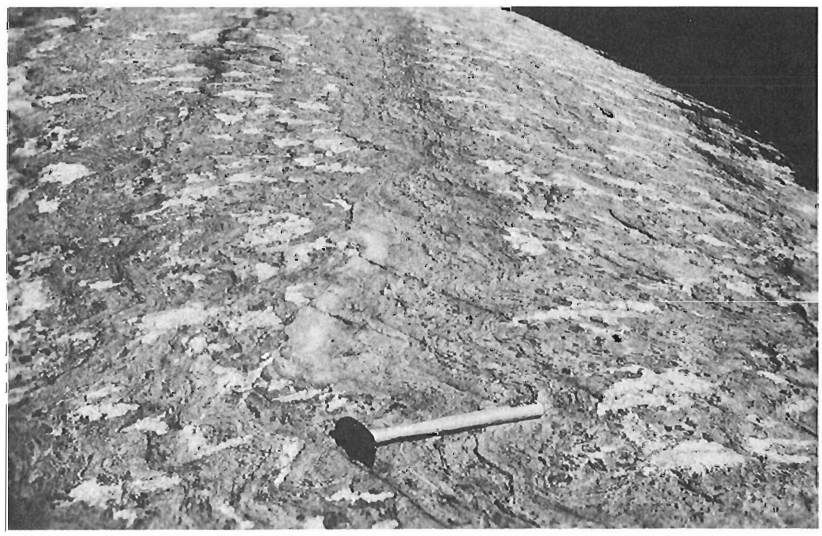


Figure 30

Selective partial melting of pelitic gneisses in interbedded sequence of pelitic and psammitic gneisses, high grade part of Kisseynew Sedimentary Gneiss Belt. (GSC 203317-C)

occur either as poikiloblasts with quartz inclusions or as discrete porphyroblasts with inclusions of anorthitic plagioclase, sillimanite, hercynite and/or magnetite. M_1 cordierite porphyroblasts in shale-derived gneisses are characterized by an abundance of acicular sillimanite crystals concentrated in their centre. These cordierite/sillimanite blasts are locally up to 18 cm long.

M_2 Metamorphism

M_2 mineral assemblages are typical of high grade (upper almandine amphibolite to hornblende granulite facies) conditions. M_2 is separated from M_1 by an episode of intense deformation, D_2 (Table 1).

M_2 caused only a slight increase in grain size in psammitic gneisses and little or no change of their M_1 mineral assemblages. The reverse is true of pelitic and semipelitic units, which contain well developed M_2 mineral assemblages. Textures in pelitic and semipelitic rocks indicate that the following reactions took place during M_2:

garnet + potassium feldspar = cordierite + biotite + quartz; (R.18)

garnet + potassium feldspar = biotite + plagioclase + quartz; (R.19)

garnet + sillimanite = cordierite + hercynite; (R.20)

M_1 hercynite = magnetite + hoegbomite[1] + corundum + sillimanite. (R.21)

M_2 plagioclase commonly occurs as blasts and extends into embayed or resorbed garnets. M_2 cordierite occurs as reaction rims around M_1 garnets and as sillimanite-free rims around M_1 cordierite/sillimanite blasts. The M_2 event is characterized by development of M_2 symplectic intergrowths replacing or partially replacing M_1 phases; for example:

M_1	M_2
Potassium feldspar + garnet	→ Cordierite-quartz symplectite + biotite
Potassium feldspar + garnet + quartz	→ Biotite-quartz symplectite or biotite-plagioclase symplectite
Potassium feldspar + plagioclase	→ Plagioclase-quartz symplectite + biotite

[1] Hoegbomite identified by X-ray.

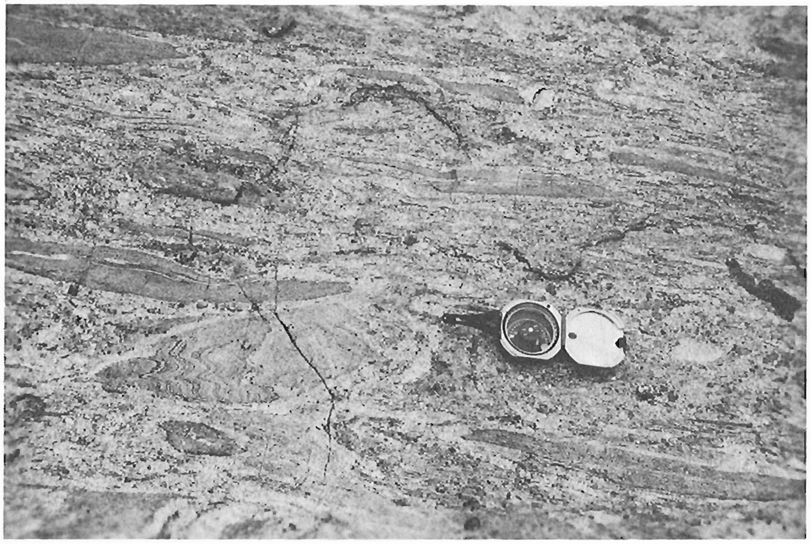


Figure 31

Unmelted rafts of psammitic gneiss floating in mobilizate derived from melting of pelitic gneisses, high grade part of Kisseynew Sedimentary Gneiss Belt. Mobilizate is rich in porphyroblasts of garnet and cordierite. (GSC 203317-A)

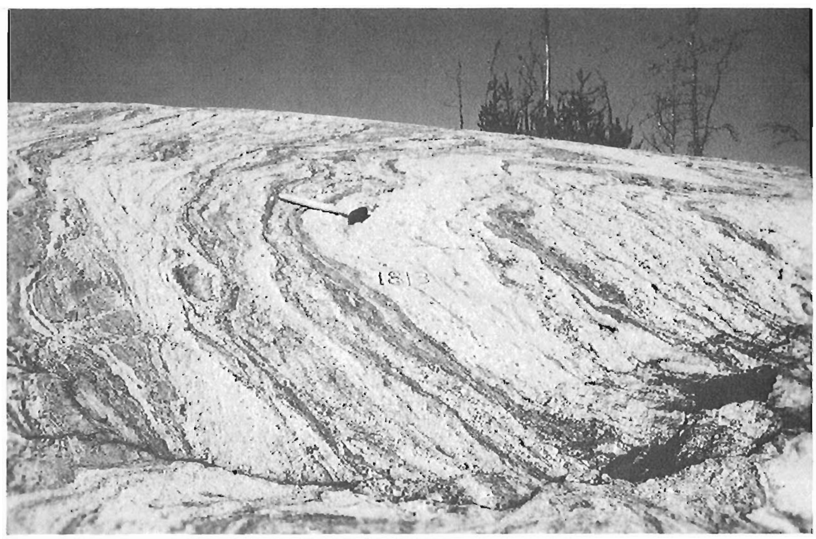


Figure 32

Lit-par-lit migmatitic gneiss with para-autocthonous white mobilizate; high grade part of Kisseynew Sedimentary Gneiss Belt. (GSC 203317-B).

M_1 hypersthene is generally stable throughout M_2, but locally breaks down to form almandine and biotite, where conditions are suitable.

ANATEXIS AND THE FORMATION OF MIGMATITES

General Statement

Bodies of white granodiorite and tonalite, which range from centimetre-wide sills to sheeted sill and dyke complexes many kilometres in extent, are characteristic of the greywacke- and shale-derived gneisses of the Kisseynew Sedimentary Gneiss Belt. It seems likely that at least some, and probably most, of these migmatitic granodiorite and tonalite complexes are products of partial melting of paragneisses.

Description of Anatectic Phenomena

A complete transition exists everywhere between pelitic and psammitic paragneisses and derived migmatites of the Kisseynew Belt. In general, there is a very good correlation between the chemistry of the original sediments and the type and amount of migmatitic product. For example, it is observed that psammitic greywacke-derived gneisses are relatively resistant to partial melting and migmatization, whereas pelitic shale-derived gneisses are very susceptible to partial melting which produces distinctive garnet and cordierite xenoblastic diatexites (Figs. 29, 30, 31).

In the File Lake area, rare quartz and quartz-feldspar segregations in pelitic rocks of the upper staurolite-biotite and lower sillimanite-biotite zone are the first observed products of anatexis. A few segregations contain small garnet and/or fibrolitic sillimanite xenoblasts. Partial melting, associated with the decomposition of muscovite, is widespread north of the sillimanite-garnet-biotite zone (Fig. 4). The granitoid mobilizate, produced by this partial melting, typically forms narrow layers in the gneisses. It is white, has a granodiorite to tonalite composition, a coarse-grained to pegmatitic texture, and generally contains garnet xenoblasts.

Migmatitic gneisses, similar to those observed in the northern part of the File Lake area, are typical of both margins (High Grade A zones, Fig. 2) of the Kisseynew Sedimentary Gneiss Belt (Fig. 32). In the centre of the belt

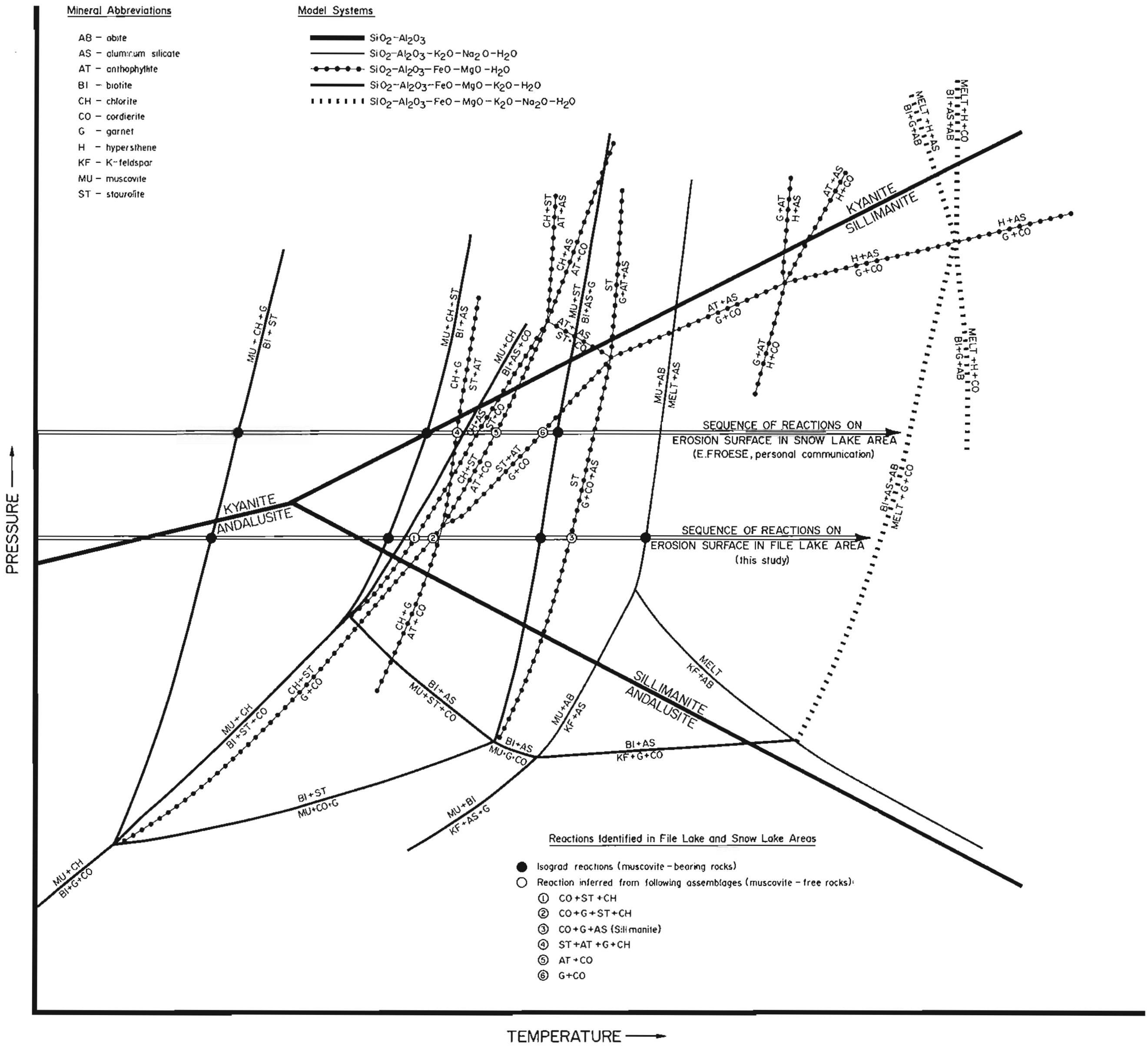


Figure 33. Petrogenetic grid (modified from D.M. Carmichael, this publication) showing reactions observed in the File Lake and Snow Lake areas, south margin of Kisseynew Sedimentary Gneiss Belt.

(High Grade B zone, Fig. 2) the partial melting of paragneisses is more complete and pervasive. This is particularly true of shale-derived gneisses, which change from garnet-sillimanite-biotite gneisses, in the High Grade A zone, to coarsely recrystallized garnet-cordierite-biotite gneisses with garnet- and cordierite-rich granitoid melts, in the High Grade B zone. The incipient melting of shale-derived gneisses is indicated by leucocratic segregations around garnet porphyroblasts. As these leucocratic segregations become more abundant, they coalesce and form lenses of leucosome, ranging from several millimetres to metres in length (Figs. 29, 30). The leucocratic segregations, which are typically rich in a residuum of garnet and cordierite, locally become mobilized and surround non-melted psammitic inclusions (Fig. 31). In essence, the process involves a concentration of lithophile elements into granitic segregations and a concentration of residual ferromagnesian elements into melt-resistant porphyroblasts of garnet and cordierite.

Interpretation in Terms of Partial Melting

Many of the migmatitic phenomena in the Kisseynew Sedimentary Gneiss Belt can be explained in terms of experimental anatexis, as summarized in Winkler (1976).

In simple terms, Winkler states that formation of a granitic melt is controlled by the reaction:

K feldspar + plagioclase + quartz + H_2O = melt (R.6)

Other than pressure and temperature, the main controls on this reaction are the availability of the reactants (i.e. K feldspar, plagioclase, quartz and H_2O). The availability of the reactants is a function of rock composition and plays an important role in controlling anatexis. Quartz and plagioclase are generally sufficiently abundant in a granitic rock for reaction (R.6) to proceed; melting is, therefore, controlled by the contents of H_2O and K feldspar. Tuttle and Bowen (1958) have shown that a lack of sufficient water will substantially increase the temperature at which melting begins; and Brown and Fyfe (1970) have shown that hydrous mineral phases, such as muscovite, biotite or hornblende, can play an important role in supplying water for the process of melting. $KAlSi_3O_8$ is an even more critical component because in its absence melting will not proceed unless very high grades of metamorphism prevail. Thus, as pointed out by Winkler (1976), in rocks above the minimum granite melting curve that do not contain primary K feldspar, any reaction which forms K feldspar component will promote melting.

In the Kisseynew Sedimentary Gneiss Belt, greywacke and shale strata do not contain primary K feldspar and, therefore, melting was controlled by reactions which generated K feldspar component. Thus, as stated in the section on prograde metamorphism in the File Lake area, the onset of melting and migmatization was controlled by the following combination of the K feldspar producing muscovite decomposition reaction (R.5) and the 'granite' melt reaction (R.6):

$$\text{muscovite + plagioclase + quartz} + H_2O = \text{sillimanite + melt.} \qquad (R.4)$$

At higher grades, further melting has occurred as a consequence of the reaction

$$\text{biotite + sillimanite + plagioclase + quartz} + H_2O = \text{cordierite + garnet + melt} \qquad (R.17)$$

which defines the boundary between the high grade A and high grade B zones in Figure 2. Reaction (R.17) represents the combination of the 'granite' melt reaction (R.6) with the reaction

$$\text{biotite + sillimanite + plagioclase + quartz} = \text{cordierite + garnet + K-Na feldspar} + H_2O \qquad (R.22)$$

The selective melting of shale-rich strata in the centre of the Kisseynew Sedimentary Gneiss Belt, and the common occurrence of anatexites rich in cordierite and almandine xenoblasts (Figs. 29, 30) is a consequence of reaction (R.17).

INFERRED P-T CONDITIONS OF METAMORPHISM

Much is known today about metamorphic reactions and the chemical composition of metamorphic minerals in pelitic schists. This information makes it possible to establish chemographic relations and construct petrogenetic grids using techniques developed by Schreinemakers (1965) and summarized by Zen (1966). A calibrated petrogenetic grid has been constructed by D.M. Carmichael (this publication) which uses a combination of experimental and field data as a basis. The metamorphic reactions observed on the erosion surface in the File Lake area (this paper) and in the Snow Lake area (Froese and Gasparrini, 1975; E. Froese, pers. comm.) are shown on an uncalibrated version of this petrogenetic grid (Fig. 33).

The sequence of reactions identified in the File Lake and Snow Lake areas correspond to those predicted at moderate pressure by Carmichael's petrogenetic grid. They indicate that a slightly lower pressure of metamorphism prevailed in the File Lake area than in the adjacent Snow Lake area.

The pressure and temperature conditions of metamorphism in the Kisseynew Sedimentary Gneiss Belt are estimated to range from 450 to 500°C, in the chlorite-biotite zone on Morton Lake, to over 750°C, in the centre of the belt. The pressure of metamorphism in the File Lake and Snow Lake areas, based on the formation of sillimanite by reaction R.2 is inferred to have been between 3.5 and 5 kb[1]. Geobarometry based on sphalerite compositions from ore deposits in the Snow Lake area indicates higher pressures, in the 6 to 8.5 kb range (Bristol, 1974; Scott, 1976). However, these apparent high pressures may be due to low temperature re-equilibration of sphalerites with monoclinic pyrrhotite; this feature has been observed in sphalerites from the Ruttan Lake Mine (C.C. Bristol, pers. comm.).

DISCUSSION AND CONCLUSIONS

Prograde metamorphic reactions identified in the File Lake-Snow Lake area are considered typical of the Aphebian high grade metamorphic event which has affected the entire Kisseynew Sedimentary Gneiss Belt, as they explain the regional distribution of minerals and mineral assemblages (Fig. 2). Anatectic phenomena observed in paragneisses of the Kisseynew Sedimentary Gneiss Belt indicate that experimental studies on anatexis of pelitic and semipelitic rocks, as summarized by Winkler (1976), are valid approximations of melting processes in natural systems. For example, melting of pelitic and semipelitic rocks in the Kisseynew Sedimentary Gneiss Belt is due to a combination of the K feldspar-forming reactions involving the decomposition of muscovite and biotite with the granitic melting reaction, as described by Winkler (1976).

Textural evidence supporting metasomatic cation exchange processes as the major mechanism of prograde metamorphic reactions (Carmichael, 1969) is abundant in the File Lake area. It seems likely that most prograde metamorphic reactions proceeded by this mechanism and that, with increasing grade, this mechanism becomes even more prominent.

In high grade metamorphic rocks in the centre of the Kisseynew Sedimentary Gneiss Belt the process of partial melting differs for different rock compositions and, consequently, the products of melting often reflect control by the original rock lithology. Thus, with careful mapping of migmatite terranes, stratigraphic units defined by character and/or amount of mobilizate can be distinguished. This is particularly true of shale-rich units which commonly give mobilizate fractions rich in refractory garnet and cordierite xenoblasts.

A feature of the Kisseynew Sedimentary Gneiss Belt, shared by many other Precambrian sedimentary gneiss belts, is that the highest grades of metamorphism are recorded in the centre of the belt and that the grade of metamorphism rapidly decreases towards the adjacent greenstone belts. Comparable relations have been described between the English River Gneissic Belt and the adjacent Bird River and Wabigoon Greenstone belts (McRitchie and Weber, 1971; Thurston and Breaks, 1978; Harris, 1976; D.L. Trueman, pers. comm.) and the Quetico Sedimentary Gneiss Belt and the Wabigoon Greenstone Belt (Kehlenbeck, 1976; Pirie and Mackasey, 1978) in the Superior Province of northwestern Ontario. In the case of the Kisseynew Sedimentary Gneiss Belt, it has been demonstrated by Byers and Dahlstrom (1954) and Bailes (1971, 1978) that the sedimentary rocks of the Kisseynew Belt and the volcanic rocks of the Flin Flon Belt are approximately contemporaneous and probably interfinger. Thus, the high grade metamorphism of the Kisseynew Sedimentary Gneiss Belt has selectively affected the gneiss belt to the exclusion of adjacent greenstone belts and probably represents an intrinsic feature of development of the sedimentary belt.

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[1] 1 kb = 1 x 10^5 kPa

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PATTERNS OF REGIONAL METAMORPHISM IN THE CHURCHILL PROVINCE OF MANITOBA (NORTH OF 58°)

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Abstract

Regional metamorphism in northern Manitoba relates to an arcuate northeast- to east-trending mobile belt 150-200 km wide flanked by regions displaying less deformation and lower grade metamorphism. Radiometric ages indicate that the sedimentary cover rocks in the mobile zone within the Wollaston domain underwent metamorphism at least 1800 Ma ago. A similar age has been inferred for metamorphism of cover rocks in the remainder of the mobile zone to the east. This age corresponds closely to the age of quartz latite of the Hurwitz Group which occurs to the northwest of the mobile zone in the flanking area of low grade metasedimentary rocks.

An earlier metamorphic-plutonic event is recorded in the granulite basement rocks between 2600 to 2700 Ma. This age was determined for basement rocks in the Wollaston domain and has been inferred for similar rocks that constitute the majority of the rocks to the east in the Nejanilini domain.

Deformation has resulted in discontinuous metamorphic gradients with the exception of the south half of the Seal River domain. In this area metamorphic grade increases from east to west, changing from greenschist to amphibolite and granulite facies metamorphism. Mineral assemblages in the zones of high grade metamorphism in the mobile belt indicate conditions of high temperature and low to intermediate pressures of metamorphism.

Résumé

Le métamorphisme régional s'est exercé suivant une zone mobile arquée, d'orientation nord-est à est, de 150 à 200 km de large, bordée au nord-ouest et au sud-est par des régions présentant des degrés moindres de déformation et de métamorphisme. Les âges radiométriques que l'on a pu déterminer indiquent que les roches de la couverture sédimentaire, situées dans la zone mobile à l'intérieur du domaine de Wollaston, ont été soumises à un épisode métamorphique, auquel on a attribué un âge minimal de 1 800 Ma. On a déduit un âge similaire pour la couverture métamorphisée du reste de la zone mobile à l'est. Cet âge correspond étroitement à celui déterminé pour une latite quartzique rencontrée au nord-ouest de la zone mobile, sur la bordure des roches métasédimentaires faiblement métamorphisées du groupe de Hurwitz.

Un épisode métamorphique et plutonique antérieur a laissé son empreinte sur les roches du soubassement granulitique, il y a 2.6 à 2.7 Ma. C'est l'âge que l'on a déterminé pour les roches du soubassement dans le domaine de Wollaston, et que l'on attribue au même type de roches, qui sont dominantes à l'est dans le domaine de Nejanilini.

La déformation a engendré des gradients métamorphiques discontinus, sauf dans la moitié sud du domaine de Seal River. Dans cette région, un gradient métamorphique, qui s'accroît d'est en ouest, se manifeste par une progression du métamorphisme, du faciès schistes verts vers le faciès amphibolite, et le faciès granulite. Dans les secteurs de degré métamorphique élevé de la zone mobile, les assemblages minéraux indiquent que le métamorphisme a opéré à une température élevée, sous une pression faible à modérée.

GENERAL GEOLOGY

The Churchill Province of Manitoba has been subdivided into domains based on lithologic and/or structural criteria. Four domains (Fig. 1) present north of 58 degrees are: the Wollaston domain, the Seal River domain, the Nejanilini domain, and the Chipewyan domain.

The Wollaston and Seal River domains have a similar lithology and metamorphism but have different tectonic histories. Both domains comprise a basement of granitoid rocks, in part hypersthene-bearing, overlain by two sequences of metamorphosed sedimentary rocks (Sequence I and Sequence II). The basement and cover rocks are intruded by igneous rocks of more than one age. The Wollaston domain displays an intense penetrative northeast-trending deformation. The structural trend of the Seal River domain is dominated by east- to southeast-trending open to tight folds, which are locally overprinted by the younger northeast structural trends. In general, the younger deformation in the Seal River domain produced open to tight minor folds and only locally formed narrow zones of intense penetrative deformation and shearing, such as the Fergus River shear zone (Fig. 1).

The Nejanilini domain is dominated by granitoid basement rocks in which thin discontinuous remnants of migmatized cover rocks are preserved. The basement rocks are in part foliated grey granodiorite to granodioritic gneiss, and in part hypersthene-bearing migmatitic rocks and foliated granitic rocks. The structural trends in this region are north of east and are intersected by numerous northerly faults.

The Chipewyan domain, an east-trending batholithic complex comprising primarily granitic to quartz monzonitic igneous rocks, is poorly exposed. The regional aeromagnetic maps indicate a broad inhomogeneity for this domain. Areas of high magnetic intensity are tentatively correlated with areas of dioritic gneiss and ortho- and clino-pyroxene-bearing rocks with igneous textures (McRitchie, 1977, and pers. comm.).

[1] Manitoba Mineral Resources Division

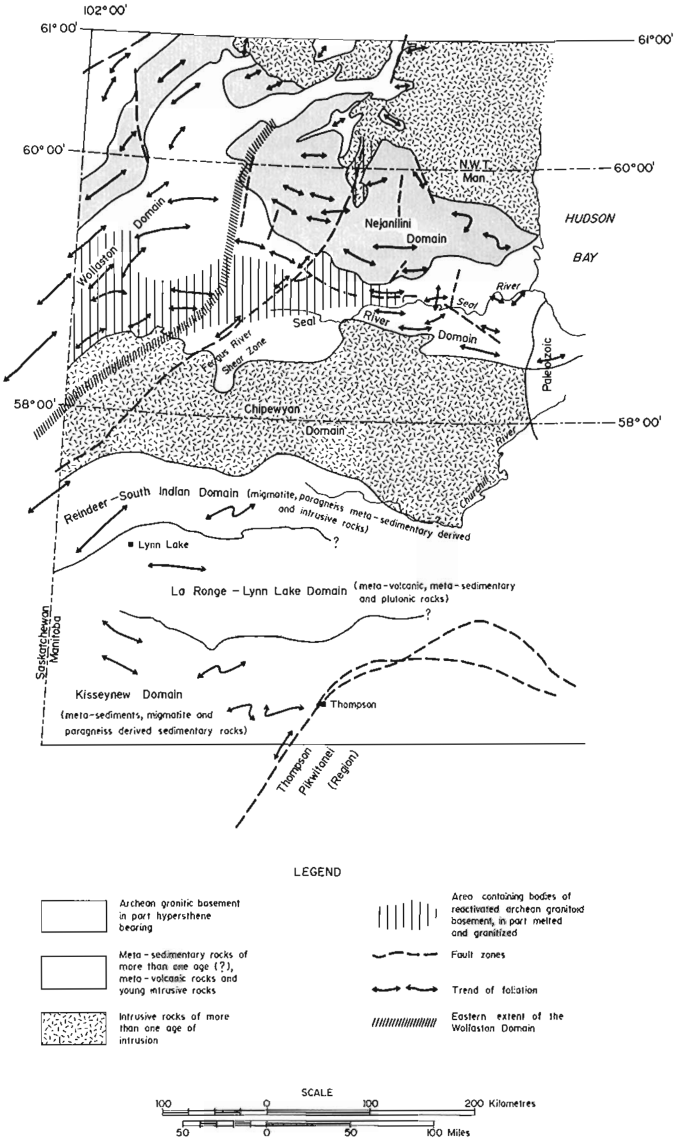


Figure 1. Tectonic domains of the Churchill Province of northern Manitoba.

Basement Rocks

Previous age studies by Money (1968) and Money and Baer (1970) indicated the presence of Archean granitic rocks in the southwest extension of the Wollaston domain in Saskatchewan. Weber et al. (1975a) documented the presence of Archean granitic and granitoid rocks. Archean ages were obtained by Weber et al. (1975b) for an intrusive suite of hypersthene-bearing rocks (2745 ± 124 Ma), and a foliated quartz monzonite (2636 ± 163 Ma).

The hypersthene-bearing suite of intrusive rocks in the southwest Wollaston domain ranges in composition from quartz monzonite to quartz diorite. The nature of the contact between these hypersthene-bearing rocks and the surrounding metamorphosed sedimentary cover rocks of Sequence I is obscured by ambiguous field relationships and the lack of adequate outcrop. A possible contact metamorphic aureole is described at one locality but conformable relationships are indicated elsewhere with no apparent contact metamorphism.

The foliated quartz monzonite, defined only in the Wollaston domain, postdates and intrudes the hypersthene-bearing rocks.

A third basement rock type, grey granodiorite gneiss, underlying the sedimentary cover of Sequence I and Sequence II in the northwest corner of the Wollaston domain in Manitoba, has yielded inconclusive age results (Weber et al., 1975a). This grey granodiorite gneiss, which displays potassium metasomatism, granitization, and intrusion on a large and small scale by young granite, yielded an age of 1941 ± 25 Ma. As indicated by Weber et al. (1975a) the grey gneiss is inhomogeneous and could contain some phases of Archean age. In the southern District of Keewatin rocks similar to the grey gneiss of the Wollaston domain in Manitoba are considered to be Archean by Eade (1971).

The complex of rocks mapped as grey gneiss and the hypersthene-bearing quartz monzonite of the Wollaston domain also occur to the east in the Nejanilini domain where they underlie the cover rocks of Sequence I. In the region between Nueltin Lake and Nejanilini Lake the grey gneiss contains lenticular bodies of weakly foliated hypersthene-bearing monzonitic rocks with intrusive textures similar to the hypersthene-bearing suite of rocks dated at the southwest end of the Wollaston domain. Once again, the contact relationships of these rocks are uncertain, but the presence of such hypersthene-bearing intrusive rocks of probable Archean age suggests that a basement complex of grey granodioritic gneiss of Archean age also underlies that part of the Wollaston and Nejanilini domains.

A grey granodioritic gneiss, which lies between Nueltin Lake and Nejanilini Lake in the Nejanilini domain, is truncated by a major north-trending fault along Nejanilini Lake. To the east of the fault the basement rocks are a complex of foliated to migmatitic monzo-charnockite and minor lenses of hypersthene- and diopside-bearing amphibolite. Small stocks of porphyritic hypersthene-quartz-monzonite are distributed throughout this terrane. The increase in the grade of metamorphism across this fault is also reflected by the mineral assemblages of the overlying rocks of Sequence I.

The basement rocks in the southeast of the Wollaston domain and the inferred basement rocks in the Seal River domain exhibit features which are indicative of intense remobilization and granitization in a highly deformed region where shearing and extension of the metasedimentary rocks has been very significant. This high level of remobilization and/or granitization of the basement rocks is maintained in a broad east-trending belt up to the west end of Great Island and appears to have been synchronous with the peak of metamorphism recorded in the Sequence I sedimentary cover rocks. Isolated inclusions of the hypersthene-bearing charnockite in the medium grained rusty-brown to reddish pink quartz-monzonite indicate derivation of this complex, at least in part, from the hypersthene-bearing basement rocks.

Cover Rocks

The cover rocks have been divided into the previously mentioned units, Sequence I and Sequence II. Rocks of both sequences are typical of a shallow water to platform environment of deposition (Table 1).

In the Wollaston domain the rocks of Sequence I contain mineral assemblages indicative of a high grade metamorphism. The basal pelitic to semi-pelitic gneiss in the south half of the Wollaston domain grades laterally to the east into a series of impure quartzite and interbedded mica schist in the southern half of the Seal River domain. The interlayered quartzite and mica schist contain mineral assemblages indicative of a low grade metamorphism. To the north of the

Table 1

Metamorphosed sedimentary cover rocks (Sequence I and II)

Wollaston Domain	Nejanilini Domain	Seal River Domain		
		West Half		East Half
		North of Seal River	South of Seal River, from Fergus River shear zone to Great Island	Great Island to Churchill
Sequence II				Sequence II
Hurwitz Group				Phyllite and very fine grained biotite schist
Metadolomite				Metagreywacke
Argillite to phyllite				Iron formation (+ magnetite + amphibole + garnet)
Metasiltstone				Metadolomite (+ quartz + clinochlore)
				Quartzite and interlayered phyllite ± garnet ± andalusite
Sheared contact zone				unconformity
Sequence I	Sequence I	Sequence I	Sequence I	Sequence I
Meta-arkose, arkosic gneiss		Meta-arkose	Quartzite, minor micaceous laminations + muscovite	
		Biotite psammite		Interlayered volcanic derived conglomerate, tuff, and metasediments
Biotite psammite gneiss	Interlayered calc-silicate rocks, marble, and biotite psammite	Calc-silicate ± marble		
Calc-silicate rocks ± marble		Quartzite ± andalusite ± diopside	Interlayered semi-pelite, quartzite and biotite muscovite schist laminations ± cordierite ± cordierite ± sillimanite ± andalusite	Metavolcanic rocks
	Semi-pelitic to pelitic gneiss			

Seal River the pelitic gneiss extends from the Wollaston domain as far east as Great Island where it interdigitates with quartzite, iron formation, and metavolcanic rocks. The mineral assemblages are characteristic of high grade metamorphism except northwest of Great Island. In this region there is an abrupt change in the metamorphic grade. The migmatitic cordierite-sillimanite-bearing semi-pelitic gneiss is truncated by an easterly shear zone and a north-trending fault zone. The grade of metamorphism in the quartzite, metavolcanic rocks, and semi-pelitic rocks east of the north-trending fault is lower amphibolite to greenschist.

The rocks of Sequence I in the north half of the Wollaston domain extend east into the Nejanilini domain where they form an irregular chain of discrete sublinear blocks. Despite the discontinuities in this chain, exposure is sufficient to indicate a gradational lithofacies pattern. In the northeast corner of the Wollaston domain, around Nueltin Lake, marble, calc-silicate rocks, and quartzite are noticeably more abundant than elsewhere in this domain. This contrast is maintained to the east in the Nejanilini domain where the basal sequence of pelitic to semi-pelitic gneiss is overlain by thickly bedded impure to clean quartzite which is sillimanite-bearing. Near the upper contact with the overlying calc-silicate rocks calc-silicate minerals are disseminated in the quartzite. The uppermost section of the Sequence I rocks contains a well-layered calc-silicate rock and biotite psammitic gneiss. Both the quartzite and calc-silicate rocks display a lateral persistence. This pattern suggests a period of emergence, standstill, and submergence. In the Nejanilini domain the meta-arkose blanket equivalent to that of the Wollaston and Seal River domains was either not deposited or was eroded. At many localities only a partial section is preserved with successively younger members of the sequence overlying the basement rocks. The rocks of Sequence I overlying the basement young to the north. This may indicate the presence of early low angle thrust faults. Younger, north to east of north-trending, steep-dipping faults are superimposed on this trend and truncate the east striking Sequence I rocks. A striking change in metamorphic grade is detectable in the cover rocks across the major north-trending fault. To the west the mineral assemblages are typical of low pressure upper amphibolite facies conditions of metamorphism, whereas to the east the assemblages are indicative of low pressure upper amphibolite to hypersthene granulite facies metamorphism.

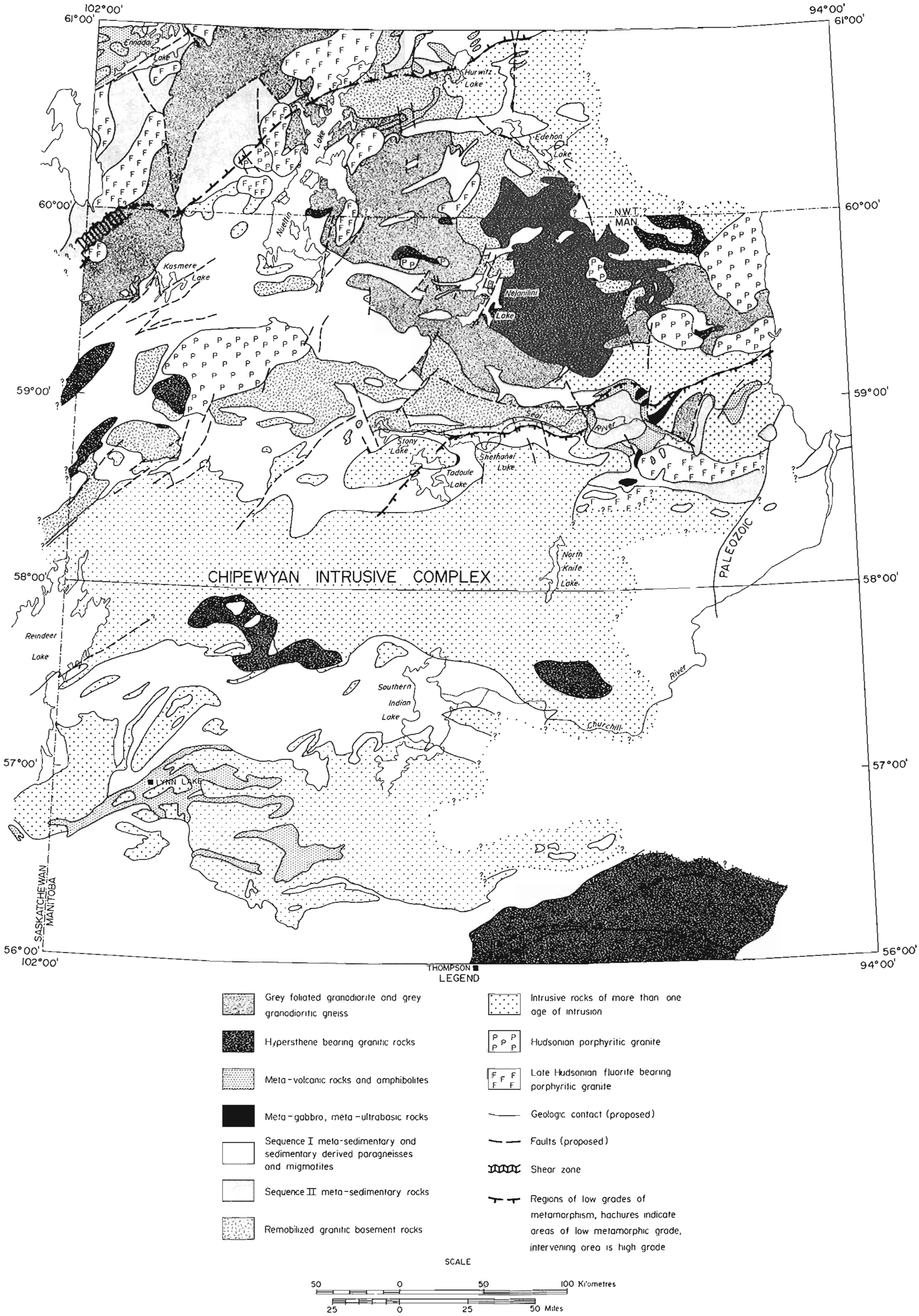


Figure 2. Simplified geologic map of the Churchill Province of northern Manitoba and the southern District of Keewatin.

The relationships exhibited between the basal pelitic gneiss and calc-silicate rocks west of the fault are maintained to the east, however, the quartzite appears to be represented by cordierite-sillimanite-quartz rocks.

The Nejanilini domain extends northerly into the southern District of Keewatin as indicated by a comparison with the geology described by Eade (1973). The sedimentary cover in the District of Keewatin, although discontinuous, exhibits many features in common with that described for the Nejanilini domain of Manitoba and may be correlative. Despite the fact that an apparent lithofacies continuity may be established a final interpretation of the relative age of the rocks in the sequence and a determination of absolute ages remains to be made.

A brief review of the inferred relative ages of the basal pelitic gneiss indicates the difficulties in attempting a correlation. Immediately north of Nueltin and Nejanilini lakes in the southern District of Keewatin, Eade suggested that the sillimanite-cordierite-garnet-bearing paragneiss, which is mineralogically comparable to the pelitic gneiss in Manitoba, is of Archean age. His interpretation is based on the assumption that the sillimanite- and cordierite-bearing paragneisses are derived from metagreywacke which interfingers with Archean metavolcanic rocks which have a minimum age of 2550 Ma (Wanless and Eade, 1975a, b). The isotopic age determinations of the pelitic gneiss in the Wollaston domain (Rb-Sr total rock isochron) yielded an age of 1800 ± 60 Ma which is similar to that obtained by Eade for the Hurwitz quartz latite near Kasba Lake. The Manitoba age has been interpreted by Weber et al. (1975a) to be the result of metamorphism during the Hudsonian orogeny, and it is thought to represent the time of isotopic closure of these fine grained rocks.

As previously stated field evidence for the contact relationships of the basal pelitic gneiss and the underlying granitoid basement rocks and hypersthene intrusive suite of rocks is ambiguous. Locally a gradational contact is indicated whereas elsewhere a sharp contact is apparent. Due to poor exposure neither contacts nor evidence of intrusive relationships were observed. Mineral assemblages in the cover rocks near contact zones are consistent with the conditions of the regional metamorphism. However, in the Wollaston domain and in the Nejanilini domain hypersthene within the cover rocks of the Sequence I pelitic gneiss is restricted to areas where the granitoid rocks contain hypersthene. These hypersthene-cordierite assemblages in the cover rocks can be explained by either: 1) contact metamorphism related to intrusion of the Archean hypersthene-bearing rocks into the pelitic gneiss; or 2) metamorphic dehydration reaction of the pelitic gneiss at the unconformable contact with a dry hypersthene granitoid basement complex. This dehydration reaction could have taken place during the remobilization of the Archean basement rocks with the resultant or attendant formation of mantled gneiss domes during the Hudsonian orogeny. It should be pointed out that the two hypotheses are not mutually exclusive. Consequently, the absolute age of the semi-pelite to pelitic gneiss remains undetermined.

Basal unconformities have been identified below shallow water sediments in both the northern corner of the Wollaston domain and far to the east on and around Great Island. In both areas the unconformity can be used to discriminate between an older Sequence I and a younger Sequence II. The Sequence II rocks are generally referred to as the Aphebian Hurwitz Group in the Wollaston domain and the Great Island Group and Churchill quartzite in the eastern area. Additional features common to Sequence II rocks in both areas include a distinctly lower metamorphic grade than the surrounding region and a spatial association with underlying volcanic rocks which are otherwise absent in the intervening areas. Specifically, in the Kognak River map area and the Hurwitz Lake areas of the southern District of Keewatin, the underlying metavolcanic rocks have been assigned a minimum age of 2550 Ma. Wanless and Eade (1975a) obtained the minimum age by dating a granodiorite which intrudes metavolcanic rocks below Sequence II rocks near Henik Lake in the Northwest Territories. In the Great Island area a similar relationship exists in which a subaerial shallow water sequence of metasedimentary rocks unconformably overlies the metavolcanic sequence, porphyritic quartz diorite to granodiorite, and volcanic derived sedimentary rocks.

Wanless and Eade (1975a) in discussing the metamorphic grade of the Hurwitz Group Aphebian metasedimentary rocks and underlying metavolcanic rocks in the Northwest Territories stated,

> "in the southern District of Keewatin there are basins little effected by orogeny and metamorphism in which both Archean and Aphebian rocks are preserved".

It can be suggested that a similar Archean-Aphebian basin existed for the Great Island-Churchill area based on the similarities of the total lithostratigraphic sequence exhibited by these two widely separated regions. This correlation would indicate that two basins now outlined by Sequence II rocks lie on the northwest and southeast flanks of a northeast- to east-trending mobile belt which records the highest grade of metamorphism and the most intense tectonic activity. The fact that this pattern is Hudsonian is established by the isotopic age, 1800 ± 60 Ma of the metamorphically high grade semi-pelitic and pelitic gneisses which contain prograde metamorphic mineral assemblages.

In the mobile zone between the two areas of low grade metamorphism, previous workers have correlated the metamorphosed sedimentary rocks which overlie the basal semi-pelite to pelitic gneisses with the Hurwitz Group (Sequence II) of the District of Keewatin (Eade, 1973). Eade interpreted the quartzite, calc-silicate-marble sequence to be not only indicative of a similar environment of deposition but also a basis for its correlation. Alternatively, workers in Manitoba (Weber et al., 1975b) equate only the uppermost unit of the sedimentary sequence, i.e., the meta-arkose and its basal biotite psammite with the Sequence II cover rocks. Unconformable relationships in the intervening mobile zone between the two distinct basins are in the form of localized unconformities at the base of the meta-arkose unit where calc-silicate cobbles occur within a quartzite and/or a biotite psammite matrix. The extent of this period of erosion and the validity of correlating it with the unconformity in the area of Great Island is uncertain. The fact that the unconformity occurs above the metapelite in the mobile zone and above possibly correlative volcanic rocks in the Great Island region supports the concept of a similar age for the unconformity. The contrasting metamorphic grades between the mobile zone and the Great Island regions may be explained on the basis of an originally different crustal setting. In such a case the Great Island Group Sequence II rocks could be of equivalent age to the metamorphically higher grade meta-arkose derived gneiss of the uppermost sedimentary rocks in the Wollaston and Nejanilini domains, and the west end of the Seal River domain.

It may be implied from this tentative correlation that in this segment of the Churchill Province an Archean crust of granitic to intermediate composition underlies a younger Aphebian suite of metamorphosed sedimentary rocks in an arcuate northeast- to east-trending mobile zone approximately 200 km wide. The mobile zone is flanked to the northwest by more weakly deformed and metamorphosed Archean rocks and is overlain unconformably by Aphebian metasedimentary rocks. A similar sequence of rocks exhibiting comparable deformation and metamorphic histories flanks the mobile zone to the southeast suggesting a similar Archean-Aphebian basin. However, the base of the Aphebian sedimentary rocks in the mobile zone and in the rocks on the southeast flank is as yet unidentified.

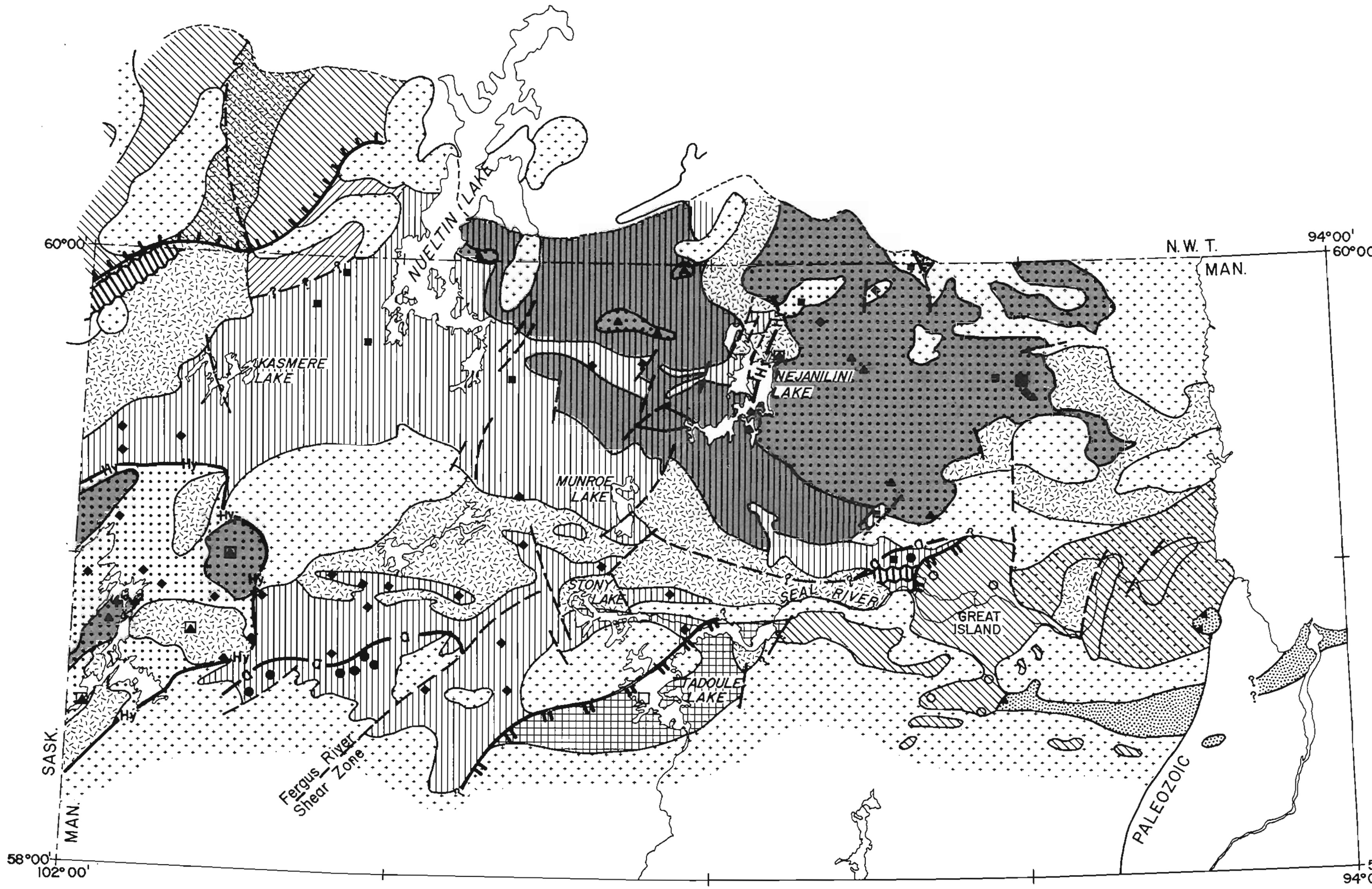
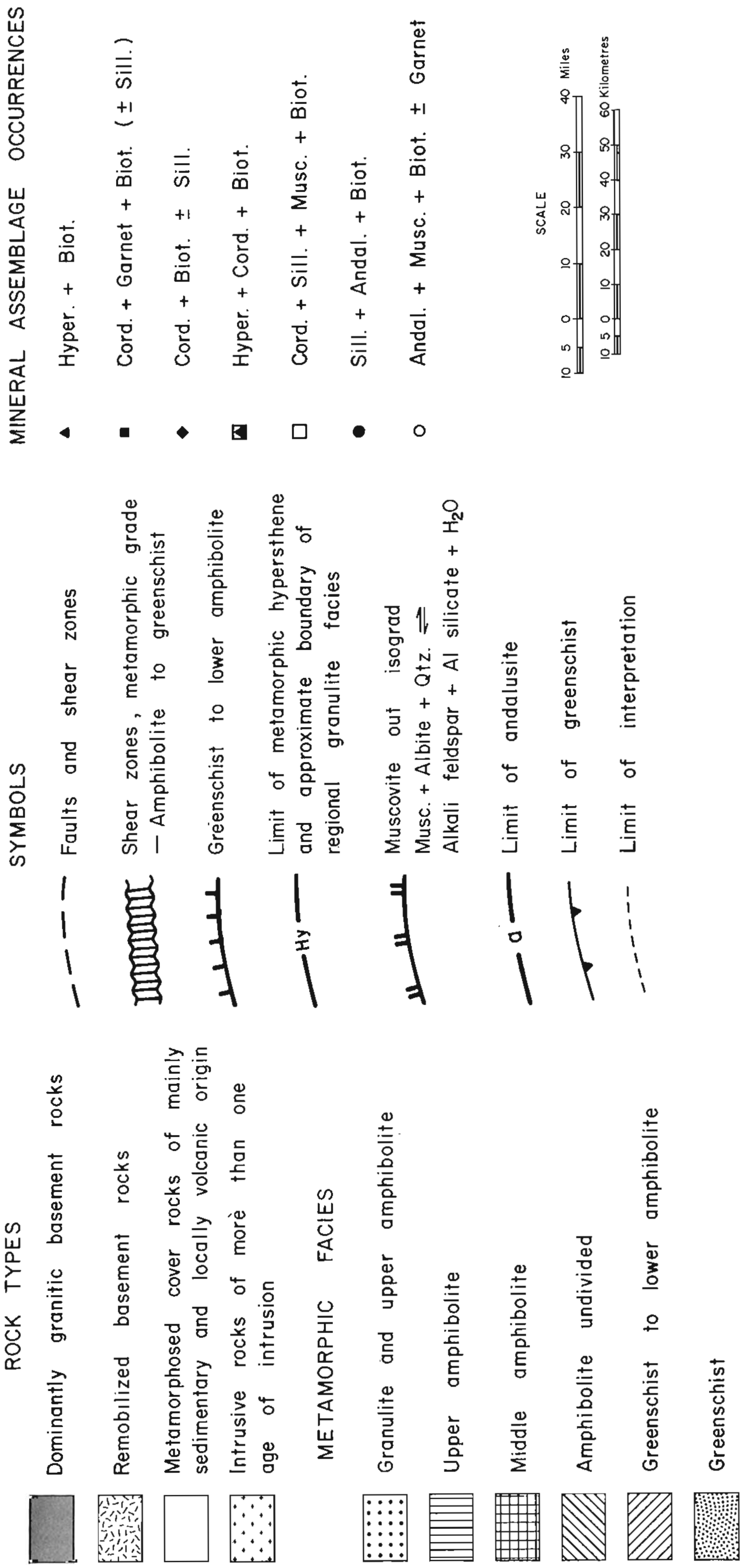


Figure 3. Simplified metamorphic facies map of the Churchill Province of Manitoba.



From the foregoing it is apparent that it is not possible at this time to demonstrate a definitive regional correlation of Sequence I metasedimentary rocks (and possibly correlative volcanic rocks) nor the younger Sequence II metasediments in the various domains. Therefore, it should be emphasized that the concept presented here of an Archean-Aphebian basin separated by the 200 km wide orogenic mobile belt of Hudsonian age is dependent on a correlation based largely on similarity of lithologics.

Mobile Zone

The mobile zone contains three distinct crustal elements. The northeast-trending Wollaston fold belt lies at the west end of the main arcuate mobile belt (Wollaston domain, Fig. 1). This zone of intense penetrative deformation has disturbed the continuity of the primary layering in much of the area on a large and small scale resulting in an almost completely pervasive northeast planar trend.

A second major crustal element is the Nejanilini Domain (Fig. 1). This rectangular block of predominantly Archean granitoid and granitic basement rocks lies in the eastern segment of the mobile belt, and appears to have behaved as a rigid body within the mobile zone. The northeastern deformational trends are only weakly developed and are superimposed on earlier east-trending folds. Younger north- and northwest-trending faults and shear zones are common in this block where a maximum crustal thickness of 50 km has been determined by Hall (1971).

A third component of the mobile zone is represented by the metamorphosed sedimentary cover that flanks the rigid Archean block, excluding the rocks of the Wollaston fold belt. The cover rocks and their underlying basement have been deformed in response to the stress field created by the interaction of the rigid block of granitic basement (Nejanilini domain) with the intense shear component of the Wollaston fold belt at the west end of the Nejanilini domain and the emplacement of the east-trending Chipewyan batholithic complex. The sedimentary cover that lies to the south of the Nejanilini massif and to the north of the Chipewyan batholith displays repeated folding and shearing about easterly trends and a major flattening of the stratigraphic sequences about these easterly trends. The underlying granitic Archean basement in this region appears to have been remobilized and deformed along with the cover rocks (Fig. 1). The late northeast structural trend, which is superimposed on these earlier structures, is developed sporadically and with variable intensity. It appears to have resulted in microscopic and minor folds and very narrow shear zones of regional dimensions. The Fergus River shear zone, which is an example of this deformational event, appears to be refracted into more north-trending fault zones where it encounters the Nejanilini domain.

The Chipewyan batholith comprises a distinct crustal element in this region of the Churchill Province and appears to have responded to the narrow zones of northeast deformation with the development of localized northeast shear zones and young north-trending faults.

METAMORPHISM

The metamorphic mineral assemblages of the Sequence I cover rocks, pelitic and semi-pelitic-derived gneiss and schist, and the aluminous interlayers within the quartzite are the most sensitive indicators of the metamorphic grade for this region. These mineral assemblages are plotted on a simplified metamorphic facies map (Fig. 3) and interpreted with respect to the megascopic and mesoscopic structural elements and the Rb-Sr total rock isochrons for the region.

The Rb-Sr total rock isochron age of 1800 ± 60 Ma has been interpreted as the time of isotopic closure of the fine to medium grained cordierite-sillimanite ± garnet-biotite plagioclase quartz gneiss from the Wollaston domain (Weber et al., 1975a). The mesoscopic structures indicate an order of recrystallization which defines the northeast structural trend of the Wollaston domain as part of the Hudsonian orogeny. Field and microscopic evidence also indicate that the deformation and metamorphism are polyphase. In the pelitic gneiss of the Wollaston domain cordierite-sillimanite and/or microcline-cordierite-sillimanite porphyroblasts are normally elongated parallel to a metamorphic layering. These porphyroblasts have been re-oriented into the northeast axial planar trend of the folded metamorphic layering. Evidence for the polyphase deformation is present in the Seal River domain where east-trending folds are sporadically intersected by narrow northeast-trending zones of intense deformation. This intersection forms either minor folds of open to tight closure or a new metamorphic layering in zones of intense penetrative deformation. As in the Wollaston domain earlier porphyroblasts are re-oriented and recrystallized without suffering retrogression. The mineral assemblage of cordierite + sillimanite ± garnet + biotite is common in the pelitic to semi-pelitic gneiss throughout the Hudsonian mobile zone (Fig. 3). This mineral assemblage and the absence of muscovite is indicative of the conditions of upper amphibolite facies of metamorphism of low pressure and high temperature. These conditions prevailed throughout most of the mobile zone during the Hudsonian orogeny.

The presence of hypersthene + cordierite + biotite ± garnet as an additional assemblage in the pelitic to semi-pelitic sequences containing cordierite + sillimanite + garnet-biotite is restricted to two regions, both of which are underlain by hypersthene-bearing granitic and granitoid rocks. At the south end of the Wollaston domain the underlying hypersthene-bearing granitic rocks have been dated as Archean. At the second locality (within the Nejanilini domain) east of the Nejanilini Lake fault, the underlying hypersthene rocks form a complex of charnockitic to charno-enderbitic gneiss, and quartz monzonite and hypersthene-bearing monzonite ± orthoclase (Davison, 1965). The granitic and granitoid rocks are thought to be Archean because they are similar to the dated Archean rocks at the south end of the Wollaston domain.

The crystalloblastic hypersthene-cordierite-biotite assemblages occur either interlayered with cordierite-sillimanite-garnet-biotite assemblages or as segregations in buff anatectic granitic layers. In the assemblage hypersthene + cordierite + sillimanite ± biotite, hypersthene, and sillimanite are incompatible, since sillimanite occurs only where it is separated from hypersthene by cordierite.

The interlayered cordierite + garnet + sillimanite + biotite assemblage, together with the co-existence of cordierite and hypersthene and the incompatibility of hypersthene and sillimanite indicates that conditions of formation for these granulite assemblages in the pelitic rocks was not significantly different from conditions of upper amphibolite facies in the rest of the mobile zone. The formation of an anhydrous mineral such as hypersthene indicates the possibility that the pelitic gneiss was dehydrated. The common denominator for both occurrences of hypersthene-cordierite gneiss is the underlying hypersthene-bearing granitic and granitoid rocks which apparently acted to reduce the partial water pressure of the overlying pelitic gneiss.

The hypersthene-cordierite assemblages occur sporadically in the pelite at the south end of Wollaston domain. This is consistent with the isolated lenticular bodies of granulite in this region. The dehydration reaction products of hypersthene-cordierite-biotite are most common in pelitic gneiss overlying the granulite of the Nejanilini massif. In this area the pelitic gneiss forms infolded, detached keels within the granulite. These small bodies of pelitic gneiss may well have been more susceptible to dehydration.

Examination of the underlying charnockitic and charno-enderbitic granulites and the hypersthene monzonite indicates localized retrograde metamorphism of the granulite related to the intrusion of Hudsonian granite. Where the alteration is intense the resultant product is a pink, biotite and/or hornblende-bearing quartz monzonite. The feldspars of this quartz monzonite generally have a variegated pink and brown colour. The layered charno-enderbitic and enderbitic granulites are altered to biotite + hornblende gneiss and amphibolite. Hybrid rocks occur as aureoles around intrusions or as isolated irregular to linear zones within the granulite complex. These isolated zones may indicate Hudsonian intrusions at depth. Portions of the rocks have relict characteristics of the granulite but no longer contain hypersthene.

Thin section examination of the granulite indicates only sporadic retrograde metamorphism. This is consistent with the preservation of Archean ages of the hypersthene-bearing intrusive rocks of the Wollaston domain. Reactions resulted in the formation of hydrous minerals. The reaction series: hypersthene → diopside → hornblende was observed in the hypersthene quartz diorite at the south end of the Wollaston domain. The reaction

hypersthene + plagioclase (An?) → biotite

is commonly observed in the foliated and layered charnockitic granulite.

There is insufficient data to definitely indicate the existence of a regional retrograde metamorphic event. Nor is there sufficient data to define or document the physical conditions during the formation of the granulite of the Nejanilini massif except to indicate that the assemblage hypersthene + garnet + biotite and hypersthene + hornblende ± biotite ± plagioclase is typical for the broad conditions of granulite facies metamorphism during the Archean.

Two zones of low grade metamorphism flank the high grade metamorphic mobile zone. Northwest of the mobile zone the metasedimentary rocks of the Hurwitz Group (Sequence II) contain mineral assemblages indicative of lower amphibolite facies metamorphism (Fig. 3). A calcareous argillite-phyllite sequence is characterized by:

plagioclase (An_{28-33}) + biotite ± tremolite.

The presence of epidote and phlogopite in a metadolomite and diopside in calc-silicate seams within the argillite suggests boundary conditions between greenschist facies and the lowermost amphibolite facies (Winkler, 1967, p. 159). The presence of cordierite in the extension of the argillite southwest into Saskatchewan and the general absence of chlorite favours lowermost amphibolite facies conditions. This region of low grade metamorphism is separated from the high grade metamorphic terrane by a zone of cataclasis several kilometres wide (Fig. 3).

The second region of low grade metamorphism occurs to the southeast of the high grade mobile zone and is outlined by the occurrence of Sequence II rocks (Figs. 2 and 3). The metamorphism of the underlying Sequence I metavolcanic rocks and the Sequence II metasedimentary rocks in the region between Great Island and the Churchill is dealt with separately.

The mineral assemblage of chlorite + epidote + albite and blue-green amphibole plus hornblende was observed within the metavolcanic rocks indicating conditions of the upper greenschist to lower amphibolite facies of regional metamorphism. Well layered amphibolite is formed within the volcanic rocks near the contacts with a pink quartz monzonite; and a porphyritic quartz diorite to granodiorite.

The metasedimentary rocks of the Great Island Group (Sequence II) overlie these two intrusive rocks and to date no contact metamorphism or any other evidence of intrusion has been observed in the Sequence II rocks where they are in contact with these igneous rocks. Present evidence suggests an unconformable relationship.

The interlayered phyllite, quartz metasiltstone and quartzite rocks of the Great Island Group display the following assemblages:

andalusite ± garnet + muscovite ± chlorite + biotite + quartz;

andalusite + muscovite + quartz;

biotite + chlorite + muscovite + feldspar + quartz, biotite as euhedral porphyroblasts.

The iron formation contains garnet and acicular amphibole while the underlying metadolomite contains accessory clinochlore and quartz. Outlying metasedimentary rocks east of Great Island and at the town of Churchill contain mineral assemblages of:

chlorite-muscovite ± biotite-quartz.

This assemblage in general indicates greenschist facies metamorphism. The general trend indicated by these assemblages in Sequence II rocks appears to be a slight lowering in the grade of metamorphism east from Great Island. The underlying contact metamorphism observed in the metavolcanic rocks has apparently been left intact by the regional lower amphibolite to greenschist facies metamorphism recorded within the metasedimentary rocks.

In addition to the slight decrease in metamorphism to the east there appears to be an increase in metamorphic grade to the south of Great Island. The Chipewyan batholithic complex lies south of Great Island. The assemblage of:

andalusite + biotite + muscovite ± garnet

occurs in very fine grained schist approximately 30 km south of Great Island. Within this area the metasediments occur as large inclusion blocks in a fluorite granite. Farther to the south the inclusion blocks become smaller and the grain size within the metasedimentary rocks increases to form coarse grained schist. The andalusite forms large porphyroblasts often folded about the hinge line of younger folds. This metamorphism is considered to be a contact metamorphism related to the intrusion of the fluorite granite.

The area of low grade metamorphism at Great Island is separated from the rocks of higher metamorphic grade of the mobile zone either by ridges or septa of intrusive rocks or faults and shear zones.

REGIONAL ZONATION

In spite of discontinuities caused by faulting and intrusion from Great Island into the mobile zone, it has been possible to define a regional metamorphic gradient. The increase in metamorphic grade appears to be continuous from Great Island to the south end of the Wollaston domain. The complete sequence of grade changes from greenschist, through amphibolite to granulite facies metamorphism is preserved only in the area south of the Seal River.

The westerly increase in the grade of metamorphism is indicated generally by changes in the metasedimentary to migmatitic rocks. Between Great Island and Tadoule Lake primary sedimentary structures are commonly preserved and are destroyed only in the noses of the tight folds, shear zones, or in regions of contact metamorphism.

Between Tadoule Lake and the Fergus River shear zone primary bedding is sporadically recognizable but has been obliterated largely by recrystallization to coarser grained schist and injection of granitic layers. Transposition of layering is more common in this region.

West of the Fergus River shear zone primary bedding is rarely preserved since recrystallization and anatexis have destroyed almost all but the largest scale variations of primary layering. Isochemical changes on a small scale in rare instances help to define original layering thus establishing a sequential development for different ages of foliation and metamorphic layering.

The mineral assemblages which define the metamorphic grade at Great Island have already been presented. The first zone to the west between Great Island and Tadoule Lake is confined to a synformal structure cored by interlayered phyllite and quartzite. The principal assemblage between Great Island and Tadoule lake is:

muscovite + biotite + quartz ± feldspar

The rocks possess a primary layering and secondary axial planar schistosity oriented north of east.

At Tadoule Lake a unique assemblage of:

cordierite + sillimanite + muscovite + biotite

occurs within a well-layered phyllite with thin interlayers of quartz metasiltstone. Cordierite forms inclusion-filled porphyroblasts 1 to 2 cm long. According to Winkler this assemblage can form only in rocks with a high Mg/Fe ratio and a low Mn content. The conditions of metamorphism are estimated at about 3 kb, and slightly higher than normal temperatures (Winkler, 1974, p. 219). The presence of fine grained groundmass muscovite distinguishes this assemblage as medium grade rather than high grade.

Between Tadoule Lake and the Fergus River shear zone muscovite is an unstable component. The muscovite isograd is indicated immediately east of Tadoule Lake (Fig. 3). However, the muscovite-out isograd is not a sharp boundary and muscovite persists up to the area of the Fergus River shear zone. It occurs as coarse to very coarse poikiloblastic grains, some containing sillimanite, as part of the "faserkiesel" comprising:

plagioclase + microcline + sillimanite + quartz ± muscovite.

This smearing out of the isograd reaction was examined by Evans and Guidotti (1966). They suggested that the divariant character of the isograd reaction is controlled by localized increases in P_{H_2O}, during dehydration, caused by conditions of low permeability. They also indicate that the poikiloblastic muscovite can be prograde while muscovite is generally breaking down. The new muscovite should theoretically contain a small amount of the sodium-mica component (paragonite) compared to higher values for the original muscovite. An alternative explanation is the flattening of the thermal gradients following the onset of partial melting.

The field and microscopic evidence in this region best fits the theory of a localized increase in P_{H_2O} during dehydration. The compatible mineral assemblage:

cordierite + sillimanite + biotite + potassium feldspar

and the field characteristics and thin section analysis suggest the loss of muscovite by the following reaction:

$$\text{muscovite} + \text{quartz} \rightarrow \text{microcline} + \text{sillimanite} + H_2O; \quad (R.1)$$

$$\text{muscovite} + \text{quartz} + \text{Na-rich plagioclase} \rightarrow \text{Na-bearing alkali feldspar} + \text{Na-poor plagioclase} + \text{sillimanite} + H_2O; \quad (R.2)$$

$$\text{muscovite} + \text{biotite} \rightarrow \text{potassium feldspar} + \text{cordierite} + H_2O. \quad (R.3)$$

The assemblage:

cordierite + sillimanite + biotite

occurs either as nebulose zones of a general planar character which coincide with primary compositional layering or as lenticular porphyroblasts parallel to or across the layering. These zones contain less biotite than the surrounding phases and are devoid of muscovite. The cordierite is of a different age than the cordierite porphyroblasts observed at Tadoule Lake in the fine grained muscovite-bearing schist. The cordierite, west of Tadoule Lake, formed at the expense of muscovite and biotite, forms interlocking and irregular poikiloblastic grains with potassium feldspar. In many instances the cordierite occurs armoured by potassium feldspar and quartz. The breakdown of muscovite takes place according to the reaction (1). The temperature interval is 580°C to 660°C and a pressure of 1 kb to 3 kb.

West of the Fergus River shear zone, the rocks are mainly migmatitic with a recognizable leucosome, melanosome, and local paleosome unaffected by migmatization. Muscovite is absent in the pelitic and semipelitic gneiss with the exception of narrow shear zones containing retrograde muscovite, epidote, and rare andalusite.

Immediately west of the Fergus River shear zone an irregular-shaped area 60 km by 10 to 20 km contains the assemblage:

andalusite + sillimanite + cordierite ± biotite.

In this assemblage sillimanite replaces andalusite (Weber et al., 1975b). The indication of this unvariant reaction:

andalusite → sillimanite (R.4)

suggests the path of metamorphism for the region passed below the aluminosilicate triple point.

Beyond the region containing andalusite the mineral assemblages:

(cordierite + sillimanite) – garnet ± sillimanite-garnet + plagioclase;

(cordierite + spinel + sillimanite + magnetite) – sillimanite-biotite-potassium feldspar

are common in the migmatite and paragneiss-derived pelite and semipelite. These assemblages characterize the upper amphibolite facies of Abukuma-type regional metamorphism.

The sillimanite occurs only as needles within cordierite and/or microcline and garnet in the presence of plagioclase. Therefore, sillimanite and biotite are unstable in the presence of plagioclase as has been reported by Reinhardt (1968). Observations for the assemblage:

biotite + sillimanite + plagioclase

indicates sillimanite occurs in plagioclase-bearing rocks if it is protected from contact with plagioclase.

As previously indicated the equilibrium assemblage:

hypersthene + cordierite + biotite

which occurs within the migmatitic-derived pelite of the southwest end of the Wollaston domain, marks incipient granulite facies metamorphism. The nature of the reactions leading to the formation of hypersthene have not been identified. The intimate relationship of the granulite facies assemblage and the assemblages of upper amphibolite facies metamorphism of low to intermediate pressure and high temperature are also indicated for the Hara Lake area of Saskatchewan. The Hara Lake area lies along the southwest extension of the pelitic and semipelitic gneiss (59° to 59°15' at 102°). Kayes (1976) has shown that Rb has not been depleted in these migmatite and gneiss in contrast to the loss of Rb which occurs in the formation of high pressure granulite.

SUMMARY OF METAMORPHIC GRADIENTS

The east- to west-trending metamorphic gradient reflects a generally linear increase with a flattening or broad levelling off in the gradient extending 50 km west of Tadoule Lake. A plateau is indicated by:

1) the persistence of metastable muscovite 30 km west of Tadoule Lake; and

2) the occurrence of the unstable pair, andalusite-sillimanite in a zone 10 to 20 km wide between Tadoule Lake and the Fergus River shear zone. This broadening of the metamorphic gradient indicates a flattening of the P-T gradient.

Farther west the isograds are closer and intermediate pressure and high temperature granulite facies metamorphism is attained in the southwest end of the Wollaston domain.

The gradient indicates that the southeast corner of the region, from Tadoule Lake to Churchill, was relatively stable compared to the mobile zone. This is consistent with the lithofacies pattern which indicates deep water to platform sedimentation in the west grading to a consistently shallow water, nearshore to subaerial stable environment followed by minor subsidence in the Great Island region during later Aphebian.

The metamorphic trend is interrupted to the south by the intrusive complex of the Chipeywan domain and to the north by the mobile zone. The tectonic elements of the mobile zone:

1) the northeast isoclinal folding and shearing followed by late northwest faults in the Wollaston domain; and

2) the east and localized block faulting in the north half of the Seal River and the Nejanilini domain,

have deformed the early regional metamorphic facies resulting in discontinuous and/or compressed metamorphic gradients.

The northern limit of the mobile zone in Manitoba is marked by a fault zone defined by a 6 km wide zone of cataclasis. The projection of the edge of the mobile zone northwest into the District of Keewatin is uncertain since it appears to bifurcate resulting in the Ennadai Lake belt of metasedimentary and metavolcanic rocks and the Hurwitz Lake belt of metasedimentary and metavolcanic rocks. A gradational contact appears to exist at the southern end of the Hurwitz Lake belt where it projects into Manitoba west of Nueltin Lake. The southern tip of the Hurwitz Lake belt of rocks has been classified as undivided amphibolite facies. The metamorphic gradient south of this projection into the mobile zone is not clearly defined and requires additional information.

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NOTES ON METAMORPHISM IN SOUTHERN DISTRICT OF KEEWATIN

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Eade, K.E., Notes on metamorphism in southern District of Keewatin; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 191-194, 1978.

Abstract

The history of metamorphism in this typical portion of the Churchill Province is complicated by the problem of the distinction of the metamorphic effects of the Kenoran and Huronian orogenies. Further, preliminary evidence suggests the presence of two additional metamorphic events in parts of the region. Post-Hudsonian metamorphism is apparently absent.

Résumé

L'histoire métamorphique de ce secteur typique de la province de Churchill est rendue complexe par la superposition des processus métamorphiques qui ont accompagné l'orogenèse de l'Hudsonien et l'orogenèse du Kénoranien. En outre, des études préliminaires indiquent que deux autres épisodes de métamorphisme auraient laissé leur empreinte sur certains secteurs de la même région. Il n'y a apparemment pas d'indices d'un métamorphisme post-hudsonien.

The area bounded by latitudes 60 to 64 degrees north and longitude 104 degrees west, east to Hudson Bay (NTS 55 and 65), for the most part in District of Keewatin but including a small part of District of Mackenzie, is in the central part of the western portion of the Churchill Province (Fig. 2). All of the problems encountered in metamorphic studies in the Churchill Province are represented in this region; Kenoran metamorphism ranging from greenschist to granulite facies with associated plutonic rocks; Hudsonian overprinting, either prograde or retrograde, with associated plutons; and possibly mild metamorphism that has affected the Helikian supracrustal rocks. In addition, evidence is beginning to appear suggesting two additional metamorphic episodes; one earlier Archean and the other mid-Aphebian.

Approximately one-half of the area has been covered by 1:250 000 mapping, a small part by 1:500 000 mapping; in the remainder the data base is from the original reconnaissance mapping published at 1:1 000 000 scale. Figure 1 shows the distribution of map areas in which more detailed information is available. The variability in the data base introduces problems in presenting a cohesive interpretation of the metamorphism in the region.

The major problem in the study of the metamorphism in this area, as in all of the Churchill Structural Province, is the distinction between the metamorphic effects of the Kenoran Orogeny and those of the younger Hudsonian Orogeny. Geochronological data in this region are limited, hence in many places lithological similarities are used to speculate on the age of rocks, and in some cases, on the age of their metamorphism. On the other hand, radioisotopic ages must also be used with care for many of the K-Ar ages indicate only the age of uplift and cooling after the Hudsonian Orogeny, and indicate neither the primary age of the rocks nor, in some cases, the age of the metamorphic minerals which they contain. It was only after dates determined by Rb-Sr and zircon methods became available that the extent of Archean terranes and the possible extent of Archean metamorphism were recognized.

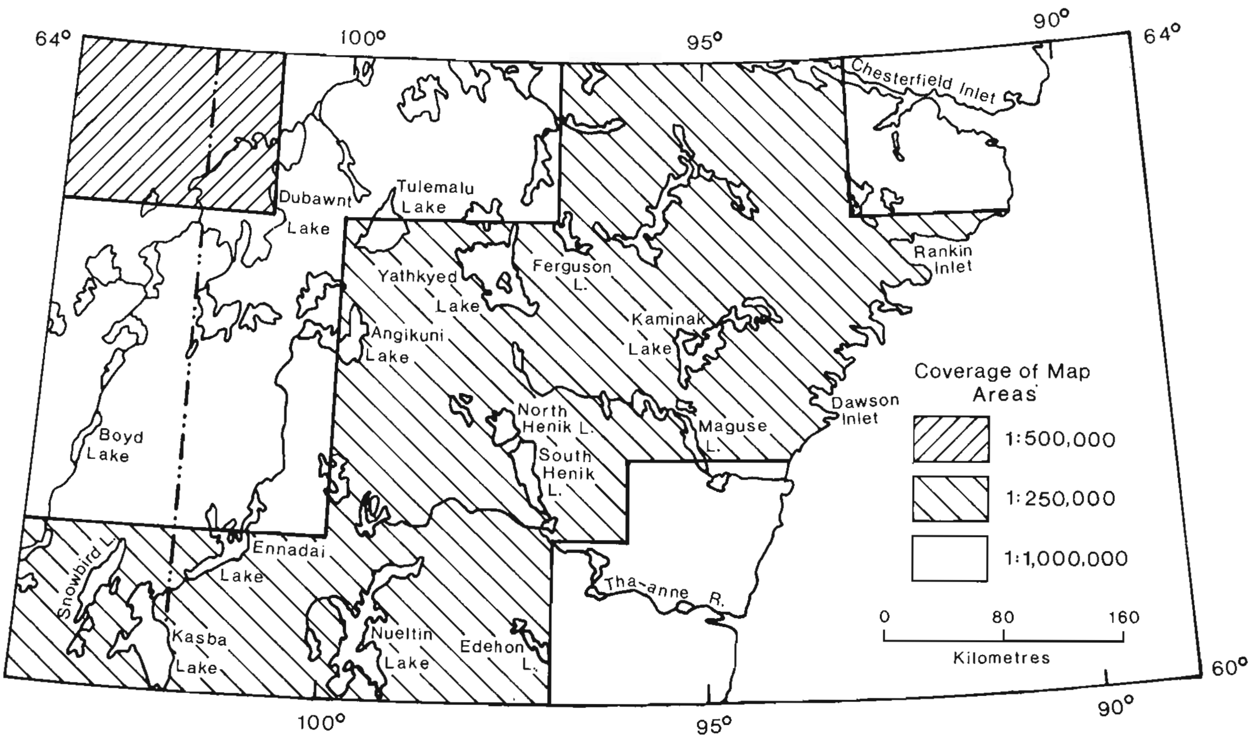
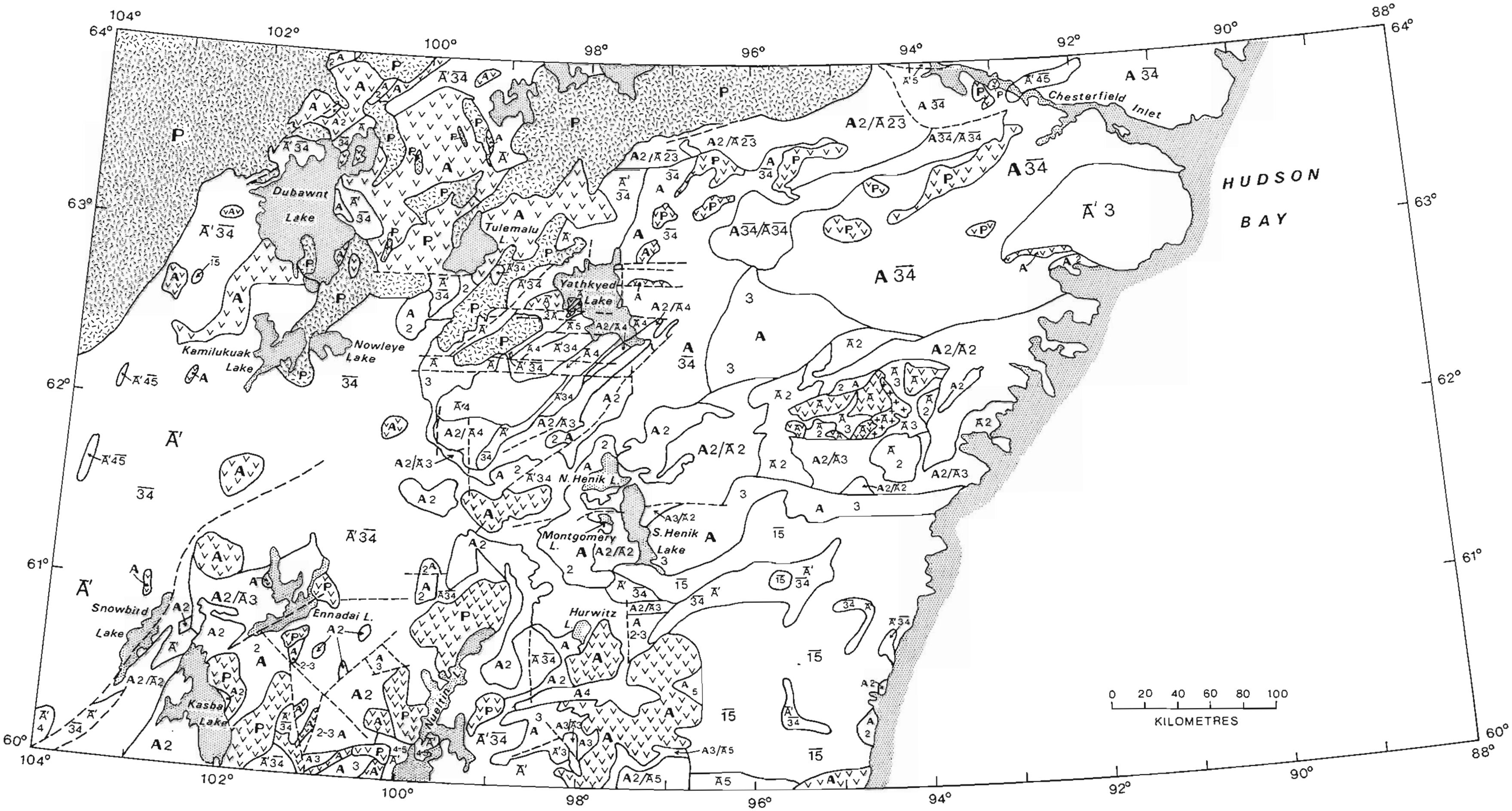


Figure 1. Geological mapping data base for metamorphic map compilation (NTS 55 and 65).

[1]Geological Survey of Canada



Age

P Paleohelikian
A Aphebian
Ā' Aphebian or Archean
Ā Archean

Non-metamorphosed rocks

Felsic plutonic rocks
Sediments and volcanics
Boundary between metamorphic map-units (approximate)
Fault
A2/Ā5 Aphebian greenschist overprinted on Archean granulite

Metamorphic Facies

5 Granulite
4 Upper amphibolite
3 Lower amphibolite
2 Greenschist
1 Subgreenschist

Undivided Facies

$\overline{45}$ Upper amphibolite-granulite
$\overline{34}$ Amphibolite undivided
$\overline{23}$ Greenschist-lower amphibolite
$\overline{15}$ Metamorphosed rocks, facies unknown

Transitional Facies

4-5 Upper amphibolite to granulite
2-3 Greenschist to lower amphibolite

Polymetamorphic Facies

$\overline{34}/\overline{34}$ Amphibolite overprinted on amphibolite
3/5 Lower amphibolite overprinted on granulite
3/3 Lower amphibolite overprinted on lower amphibolite
2/5 Greenschist overprinted on granulite
2/$\overline{45}$ Greenschist overprinted on upper amphibolite to granulite
2/4 Greenschist overprinted on upper amphibolite
2/3 Greenschist overprinted on lower amphibolite
2/$\overline{23}$ Greenschist overprinted on greenschist to lower amphibolite
2/2 Greenschist overprint on greenschist

Figure 2. Metamorphic map of the southern District of Keewatin and part of the southeastern District of Keewatin (NTS 55 and 65).

The degree of Hudsonian metamorphism can be determined, however, where Aphebian supracrustal rocks (e.g. Hurwitz Group, Montgomery Lake Group) are present. Aphebian intrusive rocks, for example abundant early Aphebian gabbro dykes cutting Archean rocks in the Kaminak Lake (Davidson, 1970a; Christie et al., 1975) and the Tulemalu Lake (Eade, 1976) map areas, provide a measure of the effects of the Hudsonian Orogeny. In some areas Archean granitic plutons of a type known to have been emplaced during the Kenoran Orogeny (Davidson, 1970a, b; Wanless and Eade, 1975) but which were unaffected by that orogeny, can also be used to estimate the intensity of the effects of the later Hudsonian Orogeny. These plutonic rocks, commonly of granodiorite to quartz monzonite composition, are not strongly affected; partial chloritization of the mafics, and turbidity of the plagioclase with development of some fine sericite, epidote and clinozoisite, is most usual. It could be argued that these effects are a deuteric alteration associated with the cooling and unroofing of the plutons and not a result of Hudsonian metamorphism. In this region, for the most part, it is assumed that Hudsonian metamorphism and not autometamorphism was responsible for this alteration.

In those areas where Hudsonian metamorphism is known, whether through metamorphism of Aphebian supracrustal rocks (Bell, 1970, 1971; Davidson, 1970a; Eade, 1973, 1974; Eade and Chandler, 1975), early Aphebian dykes, or Archean plutonic rocks, nearby Archean rocks must have been overprinted. Megascopically and microscopically it is virtually impossible, however, to separate the effects of this overprinting except where a younger cleavage is developed. Mineral assemblages commonly show nothing of this overprinting, for example, Archean metagreywacke of greenschist facies occurs adjacent to greenschist facies Aphebian metagreywacke (Eade, 1974). The Archean rocks were folded, metamorphosed, and eroded prior to deposition of Aphebian rocks. They were then overprinted by Aphebian metamorphism but no evidence can be discerned in thin section of two metamorphic events.

In extensive areas in the western part of the region where the only information available is from the reconnaissance surveys (Wright, 1967), the metamorphosed rocks for the most part are granitoid gneisses and plutonic rocks that can only be classified as undivided amphibolite facies. Although their age of metamorphism is unknown, it is thought to be largely Archean with an overprint of Aphebian metamorphism of slight to medium degree. At the present, few data are available for this region.

The southeast part of the region bordering Hudson Bay contains very few outcrops and has been mapped on only the broadest reconnaissance scale; hence metamorphic grade and age in much of this area is virtually unknown. Along the south boundary of the area narrow east-trending band of granulite facies rocks is thought to be present, the extension of a larger granulite terrane to the south. These granulites are believed to pass northward into amphibolite facies rocks of Archean age that are intruded by Aphebian granitoid plutons.

In those parts of the region where 1:250 000 mapping is completed, it is possible to make more detailed metamorphic subdivisions but even in these areas suitable metamorphic data are commonly lacking because of sparse outcrops and the absence of rocks containing diagnostic mineral assemblages. Hence isograds can be determined only in certain small areas (Davidson, 1970a; Eade, 1974). Furthermore, certain assumptions have been made in parts of these areas in determining the grade of metamorphism; for example, where migmatite occurs it is assumed that melt was present and therefore the rocks are upper amphibolite facies, although this is perhaps not always true. Small scale metamorphic features are recognized in many of the areas, for example the generally narrow contact metamorphic overprinted zones around some of the Aphebian plutons (Eade, 1973; Tella and Eade, 1978).

Plutons of Nueltin Granite are considered to be of Paleohelikian age, following Stockwell (1972), and contact metamorphic overprinting, which is associated with some of these plutons but absent from others, is considered to be Paleohelikian age. Weber et al. (1975), working in the region to the south in northern Manitoba, consider the Nueltin Granite to be of late Aphebian age as a result of their interpretation of late Aphebian – early Paleohelikian tectonic and igneous events in that region.

The Paleohelikian Dubawnt Group rocks are considered to be unmetamorphosed although they have undergone propylitic alteration (LeCheminant et al., 1977; Blake, 1978). It is possible that at least in part they have been subject to subgreenschist facies metamorphism.

Major problems of the metamorphism in this region remain unresolved at the present time, particularly in regard to the age of metamorphism in some metamorphic terranes. For example, preliminary results from an on-going dating program suggest that an orogenic event, with associated metamorphism, took place in mid-Aphebian time, i.e. between the Kenoran and Hudsonian orogenies. Such an orogenic event might account for the folding and erosion of the supracrustal rocks of the early Aphebian Montgomery Lake Group, prior to deposition of overlying later Aphebian Hurwitz Group supracrustal rocks. Although no evidence exists in the Montgomery Lake Group rocks of metamorphism accompanying the deformation (i.e. metamorphic grade is identical to that of the overlying Hurwitz Group), it is possible that elsewhere in the region metamorphism associated with this early deformation of the Montgomery Lake Group is being dated. Hence, some metamorphic terranes, presently regarded as Archean age with Aphebian age overprinting, may actually be early to middle Aphebian age, with late Aphebian age overprinting.

Elsewhere, preliminary data indicate the existence of remnants in the gneiss terranes of gneisses that are pre-3000 Ma age (R.K. Wanless, pers. comm.). These may provide evidence of an Early Archean episode of metamorphism.

It is expected that continued mapping and radioisotopic dating in this region will eventually yield hard data supporting the presence of no less than four major metamorphic events: an earlier Archean event circa 3 Ga, the Kenoran at 2.7 Ga, a mid-Aphebian event at 2.2 Ga, and finally, the Hudsonian at 1.8 Ga.

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METAMORPHISM IN THE CHURCHILL PROVINCE, DISTRICT OF MACKENZIE

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Abstract

The northwestern Churchill Province is underlain mainly by quartzofeldspathic gneiss and subordinate metavolcanic and metasedimentary rocks of Archean age which have been metamorphosed to amphibolite and granulite facies. Metamorphism to amphibolite facies probably took place during the Kenoran Orogeny. Metamorphism to granulite facies, characterized in the north, by biotite-hornblende-hypersthene assemblages, and in the southwest by garnet-cordierite-sillimanite-spinel assemblages, is possibly pre-Kenoran. Early to mid-Aphebian, greenschist to lower amphibolite facies metamorphism recognized in Aphebian strata at Great Slave Lake, may elsewhere overprint amphibolite and granulite facies rocks. Most of the Archean and Aphebian metamorphic mineral assemblages formed at low to intermediate pressures. However kyanite-staurolite assemblages, which occur only near the Slave-Churchill Province boundary and which are, therefore, probably Aphebian in age, indicate medium pressure metamorphism. The abrupt increase in metamorphic grade eastward across the northeastern segment of the Slave-Churchill boundary known as the Thelon Front is the result of uplift and erosion of Churchill terrane in Aphebian time.

Résumé

Les terrains du nord-ouest de la province de Churchill sont surtout composés de gneiss quartzo-feldspathiques accompagnés de roches métavolcaniques et métasédimentaires d'âge archéen, qui ont été métamorphisées dans le faciès amphibolite et le faciès granulite. Le premier type de métamorphisme a probablement eu lieu pendant l'orogenèse du Kénoranien. Le second, caractérisé au nord par des assemblages à biotite-hornblende-hypersthène, et au sud-ouest par des assemblages à grenat-cordiérite-sillimanite-spinelle, est probablement antérieur au Kénoranien. Un métamorphisme d'âge aphébien inférieur à moyen, qui varie du faciès schistes verts au faciès amphibolite inférieure, et que l'on a identifié dans les strates aphébiennes du Grand lac des Esclaves, a pu se superposer ailleurs au faciès amphibolite et au faciès granulite. La plupart des assemblages métamorphiques archéens et aphébiens se sont formés à des pressions faibles à intermédiaires. Cependant, les assemblages à cyanite et staurolite, que l'on rencontre seulement près de la limite entre la province des Esclaves et la province de Churchill, et qui sont donc probablement d'âge aphébien, indiquent que le métamorphisme a agi sous une pression moyenne. L'augmentation soudaine de l'intensité du métamorphisme, lorsqu'on se dirige vers l'est, après avoir franchi le segment nord-est (front de Thelon) de la limite entre la province des Esclaves et la province de Churchill, résulte du soulèvement et de l'érosion des terrains de la province de Churchill pendant l'Aphébien.

INTRODUCTION

The northwestern Churchill Province comprises a medium to high grade metamorphic terrane that presents a marked contrast against the generally lower grade terrane of the neighbouring Slave Province. The boundary between these provinces which, northeast of Great Slave Lake, is known as the Thelon Front was formerly thought to delineate the western limit of Hudsonian metamorphism (Wright, 1967). Recently the Thelon Front has been regarded as a major tectonic feature separating regions subjected to differential uplift. The origin and significance of this feature, however, are still the subjects of controversy. The metamorphic history of the northwestern Churchill Province is presumably closely related to crustal activity in the vicinity of the Thelon Front and therefore a review of metamorphic relationships in the area east of this front and in the area south of Great Slave Lake, may serve to clarify the nature of this activity.

The geology of the Churchill Province in the District of Mackenzie (Fig. 1) is known principally from the results of reconnaissance mapping. The region north of the McDonald Fault was investigated by Wright (1967), and by Fraser (1964, 1968, 1972). The region south of the fault was investigated by Stockwell (1936a, b), Henderson (1939), Wilson (1941), Brown (1950a, b, c), Wright (1951, 1952), Hoadley (1955), Taylor (1957, 1959, 1971), Mulligan and Taylor (1969), Taylor et al. (1970), and B.W. Charbonneau (pers. comm., 1978). Detailed mapping has been confined to the East Arm of Great Slave Lake (Hoffman, 1968; Hoffman et al., 1977) and to a few small areas south of Great Slave Lake (Irwin and Prusti, 1955; Reinhardt, 1969). Summaries of the geology of the Churchill Province have been prepared by McGlynn (1970) and by Davidson (1972).

This paper is based on published and unpublished data. The metamorphic mineral assemblages described are the result of examination of several hundreds of thin sections representing rocks exposed north of the McDonald Fault and also rocks exposed in the Tsu Lake area (Fig. 1). For the remaining regions, where no thin sections were available, metamorphic grade was inferred from the descriptions of map units in published maps and reports. The definitions of metamorphic facies mentioned in this paper follow Winkler (1967, 1974).

Acknowledgments

I am indebted to B.W. Charbonneau of the Geological Survey of Canada for his kindness in making available thin sections and unpublished reports covering the Tsu Lake area. The assistance of Marcia A. Mazurski who examined the Tsu Lake sections and many hundreds of sections from the northern Churchill Province collections, is gratefully acknowledged. I wish to thank P.A. Chernis for compiling data from the Armit Lake region. I also thank my colleagues in the Precambrian Subdivision of the Geological Survey of Canada, particularly W.W. Heywood, J.C. McGlynn, F.C. Taylor, and P.T. Thompson, for many informative and enjoyable discussions on the geology of the northwestern Canadian Shield.

[1] Geological Survey of Canada

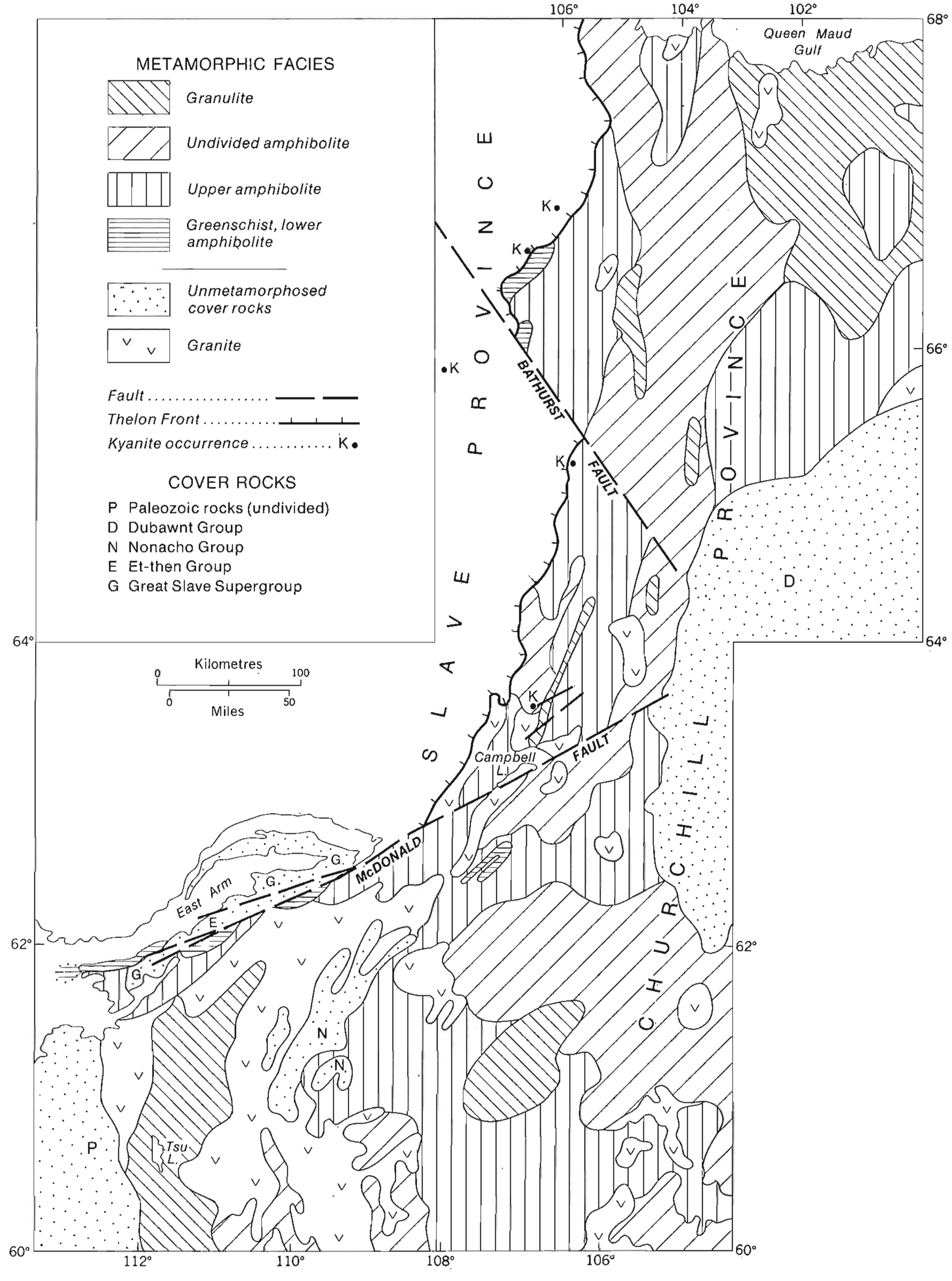

Figure 1. Distribution of metamorphic facies in the northwestern Churchill Province.

Table 1

Lithologic units and metamorphic events in the northwestern Churchill Province. Precise correlation between entries in adjacent columns is not implied.

	East Arm	Armit Lake	Queen Maud
Helikian	Mackenzie diabase dykes	Mackenzie diabase dykes	Mackenzie diabase dykes
		Northerly trending diabase dykes	Northerly trending diabase dykes
		Dubawnt Group	Dubawnt Group
	Et-then Group	Nonacho Group	
	TRANSCURRENT FAULTING	TRANSCURRENT FAULTING – MYLONITIZATION	TRANSCURRENT FAULTING – MYLONITIZATION
Aphebian	Diorite-monzonite laccoliths	Granite	Granite
	Great Slave Supergroup		
	East-northeast-trending diabase dykes	Northeast-trending diabase dykes	Northeast-trending diabase dykes
	MYLONITIZATION	MYLONITIZATION	MYLONITIZATION
	METAMORPHISM	METAMORPHISM	METAMORPHISM
	Wilson Island Group	Easterly trending mafic dykes	Easterly trending mafic dykes
Archean	Adamellite	Granite Anorthosite, gabbro	Granite Anorthosite, gabbro, norite
	METAMORPHISM	METAMORPHISM	METAMORPHISM
	Gneisses and migmatite	Tazin Group	Yellowknife Supergroup
		METAMORPHISM?	METAMORPHISM?

GEOLOGIC SETTING

The northwestern Churchill Province comprises three geologically distinct regions: Queen Maud, Armit Lake, and East Arm. The terms Queen Maud, which refers to the region north of the McDonald Fault and east of the Thelon Front (Fig. 1), and Armit Lake, which refers to the region south of the McDonald Fault, were first used by Heywood and Schau (1978) in describing geologically uniform terranes bounded by major faults that occur in the northwestern Canadian Shield. In this report the terms Queen Maud and Armit Lake are restricted, as a matter of convenience, to parts of these regions that lie within the District of Mackenzie. The sequence of formations and metamorphic episodes in these three regions is shown in Table 1.

Queen Maud Region

This region is underlain predominantly by quartzofeldspathic gneiss. Biotite and hornblende are almost invariably the chief mafic minerals but locally the gneiss is characterized by hypersthene. The hypersthene-bearing gneiss or granulite is more common in the eastern part of the region. Bodies of charnockite, anorthosite, and norite are locally associated with the granulite (Fraser, 1964). Amphibolitic schist and gneiss which may postdate the quartzofeldspathic gneiss, occur as remnants of northerly trending belts that can be traced westward into the Slave Province where they are recognized as basic metavolcanics correlative with the lava of the Archean Yellowknife Supergroup, as redefined by Henderson (1970). Quartzofeldspathic gneiss and amphibolitic rocks in general trend parallel with the Thelon Front.

Most of the granitic rocks in the Queen Maud region are probably of Kenoran age but some may be Aphebian. Potassium-argon ages obtained from the gneiss and granitic rocks range from about 1600 to 1900 Ma (Wanless, 1970). These values are considered to represent minimum estimates for ages of intrusion.

Although no exposures of Aphebian supracrustal rocks are known in the Queen Maud region, the preservation of Aphebian (Goulburn Group) strata in a down-faulted block near the Slave-Churchill boundary, 35 km east of the present Goulburn basin suggests the possibility of former Aphebian cover in this part of the Churchill Province (Fraser, 1968). Basement rocks in the eastern Queen Maud region are overlain unconformably by sandstone and conglomerate of the Helikian Dubawnt Group (Wright, 1967; Donaldson, 1965, 1967). Lava in the lower part of the Dubawnt succession exposed in the District of Keewatin to the east, has yielded a Rb-Sr age (^{87}Rb decay constant = $1.47 \times 10^{-11}y^{-1}$) of 1724 Ma (Wanless and Loveridge, 1972).

Four swarms of mafic dykes, distinguished primarily by trend direction, have been recognized. In order of increasing age these trend north-northwest, northerly, northeasterly, and easterly. The north-northwest-trending dykes belong to the Mackenzie swarm, dating about 1200 Ma (Fahrig and Jones, 1969). The northerly trending dykes, like the Mackenzie dykes are composed of fresh diabase. They have yielded K-Ar ages of 1560 Ma and 1635 Ma (Wanless et al., 1973). Most of the northeasterly trending dykes are fairly fresh but some show strain effects. Their age is uncertain. All of the easterly trending dykes have been metamorphosed and later cataclastically deformed. A K-Ar age of 1765 Ma (Wanless et al., 1973) obtained from one of these dykes is assumed to be a minimum estimate for the age of intrusion.

Armit Lake Region

The oldest rocks exposed south of the McDonald Fault consist mainly of granitic gneiss, paragneiss, calc-silicate gneiss, amphibolite, chlorite-biotite schist, biotite-hornblende schist, and minor metasediments and metavolcanics. The metasediments, unlike those of the Queen Maud region, include abundant quartzite, feldspathic quartzite, metagreywacke, and argillite. These rocks have been intruded by granite and granodiorite plutons of probable Archean age and by small bodies of gabbro and anorthosite of uncertain age.

Mulligan and Taylor (1969) correlated metamorphosed rocks of the Hill Island Lake area with those of the Tazin Group exposed in the adjacent region of northern Saskatchewan (Beck, 1969; Tremblay, 1972). On the basis of lithologic similarity most of the metamorphic rocks in the Armit Lake region should probably be included in the Tazin Group. However, some of the rocks south of the East Arm (Fig. 1) may be the metamorphosed equivalent of the Yellowknife Supergroup as suggested by Reinhardt (1969).

Although Tazin Group rocks have been variously described as Archean or Aphebian (Beck, 1969; Tremblay, 1972) radiometric age determinations leave little doubt that many of the Tazin Group gneiss and related rocks in the Armit Lake region are Archean. Thus biotite from paragneiss southeast of the East Arm yielded a K-Ar age of 2460 Ma (Lowdon, 1963). Burwash and Baadsgaard (1962) obtained K-Ar ages of 2420 Ma and 2260 Ma from boulders in Nonacho conglomerate which they considered to be Archean survival ages. Koster and Baadsgaard (1970) obtained an average K-Ar age of 2370 Ma from 20 determinations on granodiorite-diorite rocks of northern Saskatchewan. Baadsgaard and Godfrey (1972) from a dating program of rocks in northeastern Alberta that included determinations by Rb-Sr, K-Ar, and zircon methods concluded that the terrane under study is at least 2550 Ma old.

Basement rocks in the southwestern part of the region are locally unconformably overlain by strata of the Nonacho Group consisting of conglomerate, arkose, quartzite, greywacke, and slate (J.C. McGlynn, pers. comm., 1978). The Nonacho Group has been tentatively correlated with the Helikian Et-then Group (Hoffman, 1968). In the east, basement rocks are overlain unconformably by sandstone and conglomerate of the Dubawnt Group. The four dyke swarms recognized in the Queen Maud region also occur in the Armit Lake region.

East Arm Region

The East Arm of Great Slave Lake, in contrast to the Queen Maud and Armit Lake regions is underlain by thick, predominantly Aphebian, sedimentary sequences which provide a reliable record of Aphebian tectonic and metamorphic events. The oldest rocks in this region, massive Archean adamellite, granite, gneiss, and migmatite, are overlain unconformably by strata of the Wilson Island Group of probable early to mid-Aphebian age consisting of volcanic rocks and fanglomerates (Hoffman et al., 1977). Wilson Island strata are succeeded unconformably by the Great Slave Supergroup which comprises sediments and subordinate volcanics representing platform, flysch, and molasse facies deposited in an aulacogen ("Athapuscow aulacogen") associated with the Coronation Geosyncline (Hoffman, 1973). Alluvial fan sediments of the Et-then Group lie unconformably on Great Slave Supergroup strata. All rocks in the East Arm are traversed by diabase dykes of the Mackenzie swarm. The age of the Great Slave Supergroup is between 2200 Ma (Hoffman et al., 1977) and 1800 Ma (Hoffman et al., 1974).

METAMORPHISM

All the rocks in the Queen Maud and Armit Lake regions, with the exception of those in the Nonacho and Dubawnt groups and diabase dykes of northeasterly, northerly, and north-northwesterly trends, have been subjected to at least one episode of metamorphism. The grade of metamorphism increases from low amphibolite facies at the Thelon Front to granulite facies in the eastern part of the Queen Maud region, however, narrow granulite facies zones occur within amphibolite facies terrane nearer the front. No simple trend in metamorphic gradient is apparent in the Armit Lake region. In both regions facies boundaries are parallel or subparallel with boundaries of lithologic units and major structural trends. Metamorphism in the East Arm is confined to the Wilson Island Group and the underlying Archean basement. The basement has been metamorphosed to upper amphibolite facies, and the Wilson Island Group to greenschist and lower amphibolite facies (Hoffman et al., 1977). Mineral assemblages in these three regions indicate metamorphism at low to intermediate pressures, except in the vicinity of Thelon Front where andalusite-bearing assemblages and kyanite-bearing assemblages suggest that at various times either low or medium pressures prevailed.

The age of metamorphism is difficult to establish in the absence of supracrustal rocks of appropriate age. Metamorphism to upper amphibolite and granulite facies is assumed to be Archean. Lower amphibolite facies metamorphism along the Thelon Front may in part be Archean and in part Aphebian. Greenschist to lower amphibolite facies metamorphism, recorded in Aphebian strata in the East Arm (Hoffman et al., 1977) and elsewhere in the overprinting of higher grade facies, is inferred to be early to mid-Aphebian in age.

Archean Metamorphism

The interpretation of metamorphism in the northwestern Churchill Province as Archean is based in part on the previously cited K-Ar Archean survival ages that have been obtained from granite and gneiss of the Armit Lake region and adjacent regions of Saskatchewan and Alberta. It is based also on the observed continuity of Archean metavolcanic belts eastward across the Thelon Front and an assumed extension of Kenoran metamorphism into the Churchill Province. Granulite facies assemblages may derive from a metamorphic event that predates the Kenoran orogeny. This would account for the association in the Queen Maud region of mafic rocks that have attained only upper amphibolite facies with quartzofeldspathic gneiss metamorphosed to granulite facies.

Amphibolite Facies

Pelitic metasediments containing staurolite, kyanite with staurolite, and muscovite with sillimanite, associations which are diagnostic of the lower amphibolite facies, occur

Table 2

Mineral associations in granulites of the northwestern Churchill Province. Presence of mineral denoted by x.

Region	Tsu Lake																											
Quartz	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x		x
K feldspar	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x			x	x	x
Plagioclase	x	x	x	x	x	x	x	x	x	x			x	x	x	x	x	x						x	x			x
Perthite	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x			x	x	
Antiperthite				x	x																							
Muscovite	x	x	x	x	x	x	x	x	x	x	x	x	x															
Biotite	x	x	x	x	x	x	x	x	x		x	x		x	x	x	x	x	x	x	x	x	x	x	x			x
Hornblende																												x
Clinopyroxene																												x
Orthopyroxene																												x
Garnet	x	x	x	x	x	x		x		x	x	x				x	x	x		x	x	x	x	x	x			
Cordierite	x				x		x				x		x			x		x	x	x	x			x			x	
Andalusite		x	x		x	x			x					x	x	x								x				
Sillimanite	x	x			x	x			x			x	x	x			x		x	x		x	x		x	x	x	
Spinel	x	x	x	x	x		x			x	x	x	x	x	x	x			x		x	x		x	x	x	x	
Corundum	x	x	x	x										x	x									x				

Region	Thelon Front														
Quartz	x	x	x	x		x	x			x	x		x		x
K feldspar	x	x		x	x		x		x				x		x
Plagioclase	x	x	x		x	x	x	x	x	x	x	x		x	x
Perthite	x	x		x	x										x
Antiperthite						x									
Muscovite															
Biotite	x	x	x	x	x	x	x	x	x	x	x	x		x	x
Hornblende						x		x	x			x		x	
Clinopyroxene						x	x	x	x			x		x	
Orthopyroxene						x	x	x	x	x	x	x	x	x	
Garnet			x	x	x			x		x			x	x	
Cordierite	x	x	x												x
Andalusite															
Sillimanite	x	x	x	x	x										x
Spinel	x			x	x										x
Corundum	x														x

Region	Queen Maud														
Quartz	x	x	x	x	x		x	x	x		x	x		x	x
K feldspar	x	x	x				x	x	x	x	x				
Plagioclase	x	x	x	x	x	x	x	x	x		x	x	x		x
Perthite	x	x	x								x				
Antiperthite			x	x	x	x					x	x			
Muscovite															
Biotite	x	x	x	x	x	x	x	x	x	x					
Hornblende	x	x			x		x	x	x	x	x	x	x	x	x
Clinopyroxene	x					x	x		x		x				x
Orthopyroxene	x	x	x	x	x	x	x	x		x		x	x	x	
Garnet				x						x			x	x	
Cordierite															
Andalusite															
Sillimanite															
Spinel															
Corundum															

along and near the Thelon Front. The following assemblages, each of which contains quartz and plagioclase as additional minerals, are characteristic of these rocks: muscovite-biotite-garnet-staurolite; muscovite-biotite-cordierite-staurolite-sillimanite; biotite-garnet-andalusite; muscovite-biotite-andalusite-staurolite; muscovite-biotite-andalusite-sillimanite-staurolite; and muscovite-biotite-staurolite-kyanite. Similar assemblages were reported for rocks of the Slave Province near the Thelon Front in the area north of the Bathurst Fault (Thompson, 1978). It is improbable that all these assemblages formed during one Archean metamorphic event. There are only five occurrences of kyanite in the combined Slave Province and Queen Maud region, and four of these are near or on the Thelon Front. The Thelon Front is inferred to have originated in Aphebian time and the kyanite-bearing assemblages may therefore be Aphebian. Furthermore because kyanite and andalusite are stable at different pressures it is probable that the assemblages containing these minerals, although spatially associated, formed at different times. In assessing the significance of this similarity of metamorphic assemblages, it should be noted that the Slave-Churchill boundary or Thelon Front in this area is rather poorly defined. Repositioning of this boundary only a few kilometres to the southeast would result in a much greater contrast between opposing facies.

Rocks in the upper amphibolite facies are abundant throughout the Queen Maud and Armit Lake regions. In pelitic rocks this facies is defined by the stable association of sillimanite and K feldspar. Typical mineral associations observed in pelitic rocks are: quartz-plagioclase-K feldspar-biotite-sillimanite-cordierite-garnet, and quartz-plagioclase-biotite-sillimanite-garnet, and in associated mafic rocks: quartz-plagioclase-biotite-hornblende-clinopyroxene and quartz-plagioclase-biotite-hornblende-clinopyroxene-sphene. The plagioclase in this facies is commonly andesine and the biotite colour is characteristically a deep reddish brown. The presence of migmatite was used as an indicator of the upper amphibolite facies where diagnostic minerals were absent. In much of the Queen Maud and Armit Lake regions the amphibolite facies could not be subdivided. The association calcic-plagioclase with hornblende was used to distinguish amphibolite facies from greenschist facies where other criteria were lacking, and the association sphene with hornblende was similarly used to distinguish amphibolite facies from granulite facies.

Granulite Facies

Large areas underlain by rocks metamorphosed to granulite facies occur in the Tsu Lake area south of Great Slave Lake and in the area south of Queen Maud Gulf (Fig. 1). Smaller granulite masses occur east of the Thelon Front. Mineral associations in these rocks, listed in Table 2, include all minerals found together in thin section, whether in equilibrium or not. The large body of granulite in the eastern Armit Lake region is not included in the table. Data supplied by F.C. Taylor (pers. comm., 1978) indicates that this granulite is similar to that occurring near Queen Maud Gulf.

Tsu Lake Granulite

Aluminous metasedimentary and metavolcanic gneisses, associated with coarsely porphyritic granite and quartz monzonite (B.W. Charbonneau, pers. comm., 1978), extend southward from Thubun Lakes, south of the East Arm (Reinhardt, 1969) through the Tsu Lake area into northern Alberta (Godfrey and Langenberg, 1978). These gneisses are primarily characterized by the assemblage garnet-cordierite-sillimanite-spinel. Dobretsov et al. (1972) noted that the association of these minerals is diagnostic of the 'two-pyroxene' (granulite) facies. Consequently Tsu Lake rocks have been assigned to the granulite facies although orthopyroxene-bearing assemblages have been reported only in marginal parts of the Tsu Lake area (Reinhardt, 1969; Cape, 1977; Godfrey and Langenberg, 1978).

Textural relationships in the Tsu Lake granulites suggests that several of the minerals listed in Table 2 postdate the granulite facies metamorphism. Corundum, andalusite, muscovite, and chlorite invariably appear as secondary minerals but biotite, garnet and cordierite may be in part secondary. Thus spinel (hercynite?) is commonly rimmed by corundum although locally, spinel and associated corundum are together surrounded by andalusite which in turn is rimmed by muscovite and biotite. In some rocks spinel is enclosed in sillimanite or cordierite which is in part altered to andalusite and biotite. Garnet occurs as discrete crystals and also as growths on spinel. Some of the garnets are replaced by biotite and andalusite, and some by muscovite and chlorite. Fibrolite, developed in some rocks from muscovite and from biotite, must also be classified as a secondary mineral.

The association corundum-spinel-cordierite is typically stable at lower pressures and higher temperatures than the assemblage garnet-cordierite-sillimanite-spinel (Dobretsov et al., 1972) and hence may indicate contact metamorphism, possibly associated with the emplacement of the porphyritic granite that is characteristic of the Tsu Lake area. Granulites containing cordierite-sillimanite-spinel assemblages also form relatively small masses east of the Thelon Front (Fig. 1; Table 2) where they are associated with small bodies of hypersthene-bearing granulite.

Queen Maud Granulite

The granulite facies terrane that borders Queen Maud Gulf (Fig. 1) is coextensive with granulite facies rocks mapped by Heywood (1961) in the District of Keewatin to the east, and possibly also with granulites of Boothia Peninsula in the northern District of Keewatin, studied by Brown et al. (1969). Most of the granulites in the Queen Maud Gulf area contain diagnostic hypersthene. The association garnet-clinopyroxene, also diagnostic of the granulite facies but generally stable at higher pressures, was noted at one locality. Biotite, deep reddish brown in colour, and hornblende, found in most of the granulites, are assumed to be primary. However it should be noted that Brown et al. (1969) concluded that biotite and hornblende in the Boothia Peninsula granulites formed by hydration of originally anhydrous assemblages.

Aphebian Metamorphism

The occurrence of Aphebian metamorphism described earlier, in the East Arm (Hoffman et al., 1977) indicates the possibility that Aphebian metamorphism may have occurred elsewhere in this area. The amphibolite (?) facies overprinting of the Tsu Lake granulites may thus be Aphebian as suggested by Reinhardt (1969), Cape (1977), and Godfrey (1978). However there is no strong basis at present for relating the metamorphic events in the East Arm and Tsu Lake areas.

Siliceous metasediments exposed south of Campbell Lake (Fig. 1) in the eastern Armit Lake region contain the assemblage quartz-albite-K feldspar-muscovite-biotite which is typical of the greenschist facies. The age of the metamorphism is assumed to be Aphebian. The correlation of these rocks is uncertain. They are lithologically similar to the Helikian (?) Nonacho Group sediments (Fig. 1), which however are characteristically unmetamorphosed (J.C. McGlynn, pers. comm.). They also resemble rocks of the Aphebian Hurwitz Group exposed in the District of Keewatin, which have been metamorphosed to greenschist and amphibolite facies (Eade, 1971).

Pelitic metasediments near the Thelon Front, metamorphosed to amphibolite facies, contain andalusite porphyroblasts which trend northerly parallel with Archean trends exhibited by structures in the Slave Province. These porphyroblasts are overgrown by andalusite that trends northeasterly, parallel with structural trends of the Churchill Province. Nearby Goulburn Group strata of probable late Aphebian age (Hoffman, 1973) are not metamorphosed. Metamorphism of early to mid-Aphebian age is therefore inferred.

Aphebian metamorphism may well be more common in gneissic terranes in the Queen Maud and Armit Lake regions than these few examples suggest, considering the difficulty of recognizing greenschist to amphibolite facies metamorphism where it is superimposed on amphibolite grade metamorphism.

The widespread but uneven distribution of minerals stable at very low to low metamorphic grades including chlorite, muscovite, and epidote essentially represents a retrograde metamorphism to subgreenschist facies that was probably induced by changes in metamorphic conditions related to uplift and faulting in late Aphebian to early Helikian time. Regional variations in rates and magnitudes of uplift, in part reflected in variations in K-Ar values, could account for the erratic distribution of these minerals.

CATACLASIS

Most of the rocks in the Queen Maud and Armit Lake regions have undergone postmetamorphic cataclastic deformation; many are mylonitized. This deformation extends southward from Queen Maud Gulf (Fig. 1) through the Queen Maud and Armit Lake regions into Saskatchewan (Beck, 1969; Tremblay, 1972) and Alberta (Burwash and Krupicka, 1969). From studies of subsurface material from the Precambrian basement in Lake Athabasca-Great Slave Lake region Burwash (in Donaldson et al., 1976) concluded that a mobile zone formed by cataclasis of the Archean basement about 2000 Ma ago. A mid-Aphebian age for mylonitization was also proposed by Hoffman et al. (1977) based on relationships amongst Aphebian successions in the East Arm.

The mylonites possibly represent more than one age. Mylonite zones north of Campbell Lake (Fig. 1) generally parallel the Thelon Front although the distribution of these zones is irregular and may be separated by zones that are much less deformed. The northerly trending mylonites in this area are offset along east-northeast-trending transverse faults (Fraser, 1972) which are probably contemporaneous with the McDonald Fault. Mylonitized rocks south of Campbell Lake strike parallel with the McDonald Fault and may therefore represent a younger age of cataclasis. Reinhardt (1969) also reported mylonitization associated with transcurrent faults south of the East Arm. Hoffman et al. (1977), however, found no evidence in the East Arm of mylonitization associated with the McDonald Fault.

SUMMARY

The northwestern Churchill Province is underlain principally by Archean quartzofeldspathic gneiss and subordinate metavolcanic and metasedimentary rocks which, during the Archean, were metamorphosed to amphibolite and granulite facies and, during the Aphebian, were locally overprinted by greenschist and amphibolite facies assemblages.

The occurrence of Yellowknife Supergroup metavolcanic belts in the northern Queen Maud region indicates that basement schist and gneiss in this region are predominantly Archean in age. The eastward decrease in size and continuity of the belts implies that the Churchill terrane has been uplifted relative to the Slave and also that successively deeper erosion levels are exposed to the east. This interpretation is in accord with the eastward increase in metamorphic grade across the Slave-Churchill boundary and through the Queen Maud region. This metamorphism, therefore, is considered to be an extension of the Kenoran metamorphism which characterizes the Slave Province.

The association, in the eastern Queen Maud region, of quartzofeldspathic gneiss metamorphosed to granulite facies, with mafic metavolcanic rocks metamorphosed to amphibolite facies, suggests the possibility that the quartzofeldspathic gneiss or granulite represent an older basement that was metamorphosed during a pre-Kenoran event, and was subsequently uplifted and eroded before the lavas were extruded.

Aphebian metamorphism is most clearly demonstrated in the East Arm of Great Slave Lake where supracrustal sequences of Aphebian and Helikian age occur. In the Queen Maud and Armit Lake regions, on the other hand, supracrustal rocks of Aphebian age are lacking and the effects of Aphebian metamorphism are not readily distinguishable from those of Archean age. Consequently, in these regions the extent of Aphebian metamorphic events may be seriously underestimated.

Metamorphic mineral assemblages in the northwestern Churchill Province indicate metamorphism at low to intermediate pressures. Kyanite-bearing assemblages, however, which are restricted in occurrence to the vicinity of the Thelon Front, indicate metamorphism at medium pressures. The kyanite is probably of Aphebian age, whereas spatially associated andalusite may be Archean or Aphebian in age.

The widespread, retrograde development of chlorite, muscovite, and epidote is thought to be related in part to the uplift that affected the northwestern Churchill Province in the Aphebian Era. The timing of uplift is uncertain. In the Queen Maud region uplift probably dates from early Aphebian and provided a source of material for the late Aphebian Coronation Geosyncline (Hoffman, 1973). The Armit Lake region, where possible Aphebian deposits are preserved, may have stabilized at an earlier date, as implied by Heywood and Schau (1978). Aphebian (Hurwitz Group) sediments, metamorphosed to greenschist facies (Eade, 1971) occur in the eastern extension of this region.

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METAMORPHISM OF THE PRINCE ALBERT GROUP, DISTRICT OF KEEWATIN

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Abstract

In the District of Keewatin the Prince Albert Group consists of northeast-trending belts of metamorphosed Archean quartzose, argillaceous and/or calcareous psammite and pelite interbedded with magnesian volcanic flows and iron-rich sediments.

In detail, pelitic and arenitic rocks in the western region are characterized by kyanite-chlorite and chloritoid-chlorite, garnet-chlorite-biotite assemblages, in addition to quartz and muscovite. These are characteristic of medium pressure, low temperature metamorphic environments. The central region contains cordierite-garnet, cordierite-staurolite, and rarely, andalusite-biotite, as well as quartz and muscovite. These assemblages are commonly found in medium temperature, low to moderate pressure metamorphic regimes. Rocks in the eastern region contain sillimanite, K-feldspar, muscovite, and quartz with biotite and/or garnet. These are high temperature assemblages formed at low to moderate pressures. This sequence suggests that during the peak of metamorphism temperatures were higher in the northeast than the southwest. The lower temperature rocks in the western region were apparently recrystallized at higher pressures than the higher grade ones.

The age of peak metamorphism is not proven, but can be argued to be Archean. The K-Ar radiometric data suggest local mid-Aphebian and extensive "Hudsonian" recrystallization that may correspond to a later greenschist grade diaphthoresis associated with widespread northeasterly to easterly trending faulting.

Certain implications result from the occurrence of moderate pressure assemblages. The Prince Albert Group is not thought to be thick enough to account for the depths required by the metamorphic conditions. Extensive flow folding or perhaps nappes sliding from the higher grade area may have provided the thicknesses required. The calibration of the petrogenetic grid will govern which of many tectonic models is utilized. A model favouring a thick Archean crust, and highly variable geothermal gradients, was chosen.

Résumé

Dans le district de Keewatin, le groupe de Prince-Albert comprend des zones orientées nord-est de pélites et psammites métamorphisées, d'âge archéen; ces dernières sont siliceuses, argileuses ou calcaires, et interstratifiées avec des coulées de lave magnésienne et des sédiments ferrifères.

Dans la région ouest, les roches pélitiques et arénitiques sont caractérisées par les assemblages cyanite-chlorite, chloritoïde-chlorite, et grenat-chlorite-biotite, outre le quartz et la muscovite. Ces assemblages sont caractéristiques des milieux qui ont été soumis au métamorphisme sous une pression moyenne, à basse température. La région centrale continent les assemblages cordiérite-grenat, cordiérite-staurolite, et parfois andalusite-biotite, ainsi que du quartz et de la muscovite. On rencontre généralement ces assemblages dans les milieux qui ont été soumis au métamorphisme à une température moyenne sous une pression faible à modérée. Dans la région est, les roches contiennent de la sillimanite, du K-feldspath, de la muscovite et du quartz, ainsi que de la biotite ou du grenat, ou les deux à la fois. Ces assemblages se sont formés à une température élevée, sous des pressions faibles à modérées. Cette progression de l'intensité du métamorphisme d'ouest en est indique sans doute que pendant la culmination du métamorphisme, les températures ont été plus élevées au nord-est qu'au sud-ouest. Dans la région ouest, les roches soumises à des températures moindres ont apparemment recristallisé à des pressions plus fortes que les roches de degré métamorphique plus élevé.

On n'a pu déterminer avec certitude l'époque de l'apogée du métamorphisme, mais on suppose qu'elle se situe à l'Archéen. Les données radiométriques K-Ar indiquent localement une phase de recristallisation d'âge aphébien, et ailleurs, une phase de recristallisation datant de l'"Hudsonien", qui correspond sans doute à un métamorphisme rétrograde tardif, d'intensité équivalente à celle du faciès schistes verts, et associé à la formation à grande échelle de failles d'orientation nord-est à est.

On peut tirer certaines conclusions de l'existence d'assemblages formés sous une pression modérée. On ne pense pas que le groupe de Prince-Albert ait atteint une épaisseur suffisante pour permettre le développement du métamorphisme considéré. D'importants plissements isopaques ou même le glissement de nappes à partir de la zone de degré métamorphique plus élevé ont pu créer l'épaisseur favorable aux conditions du métamorphisme. L'étalonnage de la grille pétrogénétique permet de faire un choix parmi les nombreux modèles tectoniques. On a choisi un modèle comportant une croûte archéenne épaisse et des gradients géothermiques très variables.

[1]Geological Survey of Canada

INTRODUCTION

The study of metamorphic grade in granitoid gneiss terranes is commonly difficult because minerals that define particular pressure and temperature ranges are rare. However, sedimentary differentiation in supracrustal sequences provides sufficient variety in bulk compositions to permit development of metamorphically significant mineral assemblages. The estimates of metamorphic grade for supracrustal rocks can be considered as the minimum grades attained by the associated and older gneiss.

Reconnaissance studies in the northern part of the District of Keewatin (Heywood, 1961, 1967) have shown that there are vast tracts of granitoid gneiss in the Churchill Province, which contain remnants of Archean supracrustal rocks (Davidson, 1972; Wanless and Eade, 1975) as well as later, less deformed, Aphebian supracrustal rocks such as the Chantrey Group (Heywood, 1961; Davidson, 1972). Granitoid gneiss underlies about 65 per cent of the region and amphibolite, amphibolite gneisses, greenschist and minor sedimentary schist and gneiss, underlie 25 per cent of the region (Eade and Fahrig, 1971). About 10 per cent of the area is underlain by massive, locally foliated and/or porphyritic granite to granodiorite. Less than one per cent of the area consists of local, very siliceous units. The Prince Albert Group, approximately 20 per cent of the area (Fig. 1), consists of siliceous sediments, associated metasedimentary units, and interbedded amphibolite. Since completion of the reconnaissance work in the District of Keewatin, the Prince Albert Group has been mapped in greater detail (Campbell, 1973, 1974; Frisch, 1973, Schau, 1973, 1974, 1975a, 1977).

PRINCE ALBERT GROUP

The Prince Albert Group in the District of Keewatin (Fig. 2) is distributed in northeast-trending belts. These belts, as much as 20 km wide, narrow eastward and extend for at least 280 km along strike. The lithologies change along strike from siltstone, phyllite, quartzite (Schau, in prep.), silicate, oxide and sulphide iron-formation, serpentinized ultramafic sills, dykes or flows (Schau, 1975a) and rare carbonate in the southwest, to paragneiss, sillimanite-knotted quartzite, lean iron-formations and ultramafic lenses in the northeast (Campbell, 1974). Tight, subhorizontally-plunging northeast-trending folds with variable but generally steep axial planes in the west, and generally gently inclined axial planes in the east, are crossed by later, minor southeast-trending fold sets. The trend of the belts is parallel to the trace of the beds and axial planes of the northeast-trending folds. Individual lithologic units are difficult to trace because of poor exposure, but magnetite-quartz iron-formations can be traced on aeromagnetic maps through covered areas with considerable success. Hill-forming quartzite units up to 200 m thick parallel these magnetic anomalies and may represent a more or less continuous, albeit possibly diachronous, sequence. The three schematic sections (Fig. 3) show that the upper part of the Prince Albert Group is better represented in Area A (Fig. 1). The estimated minimum thickness of the section is about 1.9 km. Individual units vary in thickness and composition along strike.

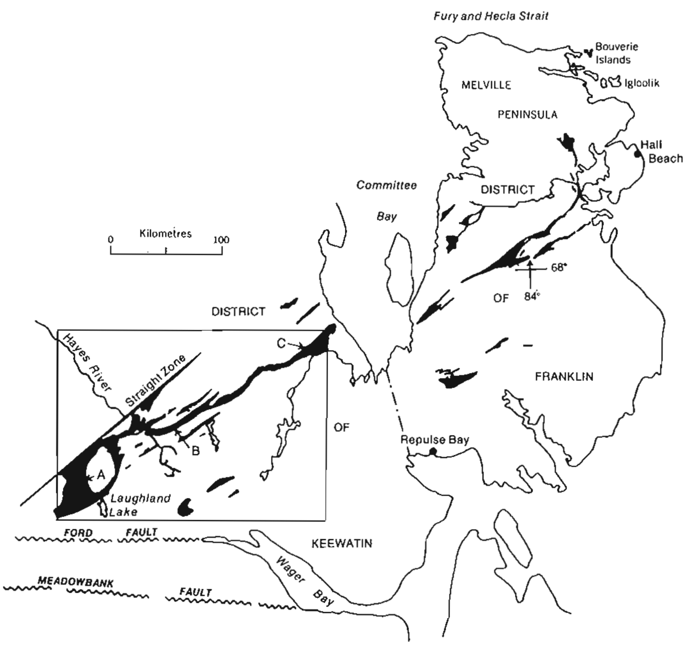


Figure 1. General location of the Prince Albert Group, Districts of Keewatin and Franklin. Outlined area is discussed in text and shown on Figure 2. Smaller areas discussed in text are labelled A, B, C.

The Prince Albert Group shows differing relations along the belt, with the bounding granitoid gneiss. In the west, faults, conformable contacts, and one possible unconformity (Schau, 1974) separate the supracrustal rocks from the gneiss. In the east, granodioritic gneiss intrudes or is formed from the Prince Albert Group (Frisch, 1973; Campbell, 1973, 1974). Plutons of different ages cut the group. Porphyritic, fluorite-bearing plutons of late Aphebian or early Helikian age cut the group near Wager Bay. Red-stained biotite granite stocks in the central part of the region may have caused local metamorphism. These stocks postdate diabase dykes that are at least as old as mid-Aphebian. Plutons in the central part of the mapped region, older than these dykes, are muscovite and garnet-bearing. Massive to foliated intrusive granodiorites occur throughout, but their ages are not well known. In the east, thick sheets of white muscovite-bearing pegmatite are interleaved with the Prince Albert Group and pink microcline-rich pegmatite extends from the bordering gneiss into the group. In the west, pegmatite dykes are less abundant and cross the regional structures near granitoid stocks. Most of the gabbro stocks that occur in the western region were emplaced prior to the main deformation, commonly near ultramafic flows, dykes, or sills, but some stocks postdate this deformation.

The age of deposition of the Prince Albert Group is Archean. A zircon from an acid volcanic unit of the Prince Albert Group on the west side of Melville Peninsula is at least 2728 Ma on the basis of a discordant $^{207}Pb/^{208}Pb$ age of zircons (Wanless, pers. comm.). Correlation of the Prince Albert Group on Melville Peninsula and the current area of interest was made by the first workers in the region (Heywood, 1967). Supporting this correlation across Committee Bay are the similarities in structural trends and unusual lithologies that include quartzite with chrome muscovite and magnesian volcanics with spinifex textures (Eckstrand, 1975; Frisch and Jenner, 1975; Schau, 1975a, 1977).

METAMORPHISM

The metamorphism of the Prince Albert Group and the "old" gneiss may be used to explain in part the variation in structural complexity, different amounts of pegmatite along strike, and the nature of contacts with the bounding gneiss. The metamorphic grade of the Prince Albert Group (Fig. 2), according to the definitions used in this symposium, ranges from greenschist to upper amphibolite. In pelitic and aluminous arenitic rocks biotite and garnet are present throughout the group. In the southwest upper greenschist assemblages are found; to the northeast, kyanite and chloritoid disappear; staurolite and andalusite are formed, and are followed by sillimanite-muscovite and sillimanite-muscovite – K-feldspar. At Committee Bay a decrease in metamorphic grade is evident from staurolite-sillimanite-quartz-muscovite associations. Pegmatite is especially abundant in the sillimanite-muscovite – K-feldspar-rich region. Although the relative positions of isograds are certain, their detailed loci are not known with great precision. It is clear, however, that they transect the main lithologic units and fold structures which in the west are tight and upright but to the northeast become isoclinal and dip less steeply.

Granitoid gneiss in the east and west contains mineral assemblages compatible with amphibolite grade metamorphism (Schau, 1975b; Frisch, 1973). Thus in the east gneiss and supracrustal rocks are of the same grade, whereas in the west the bounding gneiss is of higher grade than that of the Prince Albert Group (Schau, 1975b).

Retrograde greenschist metamorphism is sporadically and partially overprinted on granitoid gneiss and supracrustal rocks of all grades as indicated by epidote-chlorite and sericite-biotite-chlorite mineral assemblages.

The age of metamorphism is uncertain, but is probably Archean. Structural information suggests that the metamorphism and deformation occurred prior to the intrusion of mid-Aphebian gabbroic dykes. The assignment of this main deformation and metamorphism to the Archean is deduced from an analagous situation on Melville Peninsula where strongly gneissose rocks of the Prince Albert Group are known to be of late Archean age (2750 Ma, R.K. Wanless, pers. comm.). Here gneiss derived from the Prince Albert Group yields U-Pb ages on zircons of 2605 Ma (R.K. Wanless, pers. comm.). The suggestion that some northeast-trending folds of the Churchill Structural Province are Archean is deduced from an analagous situation. The suggestion that some northeast-trending folds of the Churchill Structural Province are of Archean age has previously been made by Eade and Blake (1977) for the rocks in the Angikuni Lake area. There as well as in the area under consideration, locally recrystallized mid-Aphebian dykes confirm that the region was subjected to brittle deformation. K-Ar ages obtained on platy minerals are thought to reflect degradation of the previously metamorphosed and deformed rocks associated with widespread faulting which led to the greenschist overprint, rather than to represent some initial metamorphic age.

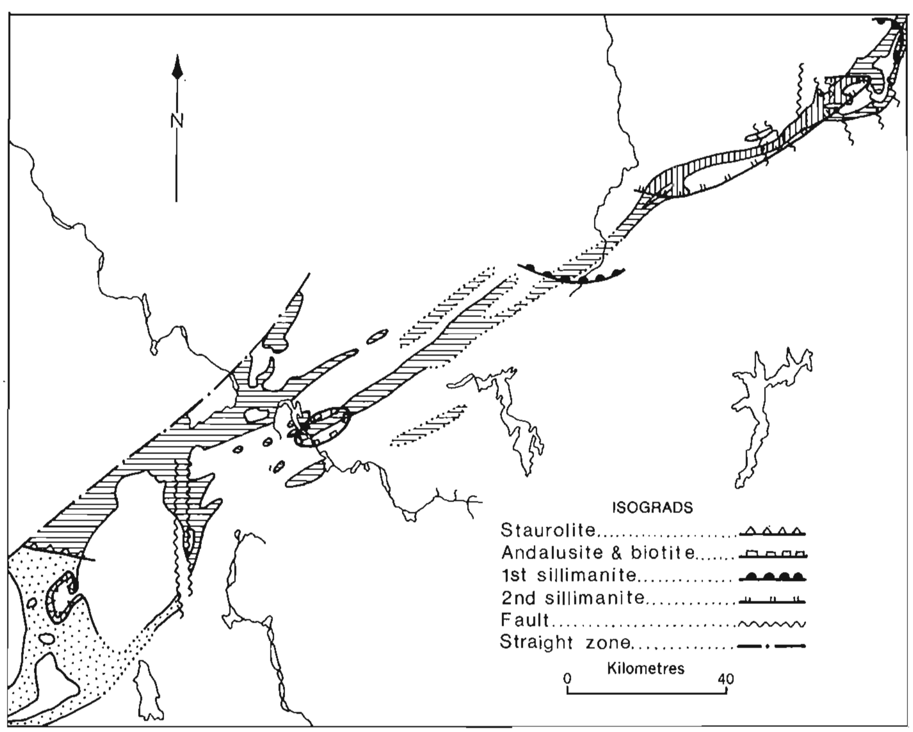
A Matter of Pressure

The general pattern of the succession of the mineral zones indicates that temperature increased to the northeast to a culmination near the border of the area shown in Figure 2. Three areas (Fig. 1) one from each of the regions where aluminosilicate minerals occur were studied in somewhat greater detail to resolve whether or not the regions were metamorphosed at the same pressure.

The uncalibrated petrogenetic grid of Ermanovics and Froese (1978) has been used in this paper to position the various mineral assemblages in P-T space. Published results of experimental work bearing on each assemblage will be reviewed to confirm the relative positions on this grid and the consequences of choosing either of two different calibrations will be presented.

Region A – Greenschist Region

Prince Albert Group strata in this region (Fig. 4) have been metamorphosed to greenschist facies grade, the lowest grade represented in the group. These strata are in both conformable and fault contact with the bordering basement(?) gneiss. In region A the group consists of phyllite, aluminous iron-formation, siltstone, quartzite containing varying amounts of alumina, and mafic sills, dykes or flows, commonly quite magnesian and probably of komatiitic heritage. Gabbro stocks and quartz-diorite to granodiorite plugs have intruded the supracrustal rocks. The major structure appears to be a large upright synform with shallow plunges to the east that deform previous tight to isoclinal folds. Metamorphic grade within the group increases gently towards the north and abruptly towards a small crosscutting granitoid pluton.



= greenschist facies
= lower amphibolite facies
= upper amphibolite facies

Figure 2

Metamorphism of Prince Albert Group.

SECTIONS OF PRINCE ALBERT GROUP		
A	B	C*
Intrusive Contact		
Magnesian Volcanic Flows(?)		
Greenstone		
Chlorite schist		
Carbonate (200 m)		
1100 m – Greenstone		Biotite Schist
Iron-Formation	Complex Structures	
Muscovite Phyllite	Pelitic Schist	Metagreywacke
White Quartzite	White Quartzite (200 m)	White Quartzite
Muscovite Phyllite	Biotite-Schist+Psammite	
	Garnet Schist	Garnet Sillimanite Schist
Iron-Formation..........................800 m	– Iron-Formation..................	.Iron-Formation
Complex Structures	Pelitic Schist	Biotite Schist
	Magnesian Volcanics	Metagreywacke
	Contact Uncertain	Biotite Schist
	Gneiss	

*After Campbell 1973 (Structural Repetition and Tops Uncertain)

Figure 3. Simplified composite sections of the Prince Albert Group from each of regions A, B, and C.

Quartzite units contain muscovite, chromian muscovite, kyanite, and accessory minerals which include tourmaline. Mineral assemblages, from rocks of pelitic composition, representing most of the area, are plotted on Diagram A1 (Fig. 5), a Thompson-type AFM diagram (Thompson, 1957). The assemblages kyanite-chlorite-quartz-muscovite ± chromian muscovite, chloritoid-chlorite-quartz-muscovite ± chromian muscovite, and garnet-chlorite-biotite-quartz-muscovite, indicate that these rocks are in the upper greenschist facies. These assemblages lie in region A1 on the uncalibrated petrogenetic grid (Fig. 6).

The higher grade rocks, on the north side of Region A, contain staurolite-biotite-quartz-muscovite and garnet-biotite-K feldspar-quartz-muscovite assemblages (Diagram A2, Fig. 5). Kyanite-muscovite-quartz assemblages with rare garnets which occur in the metamorphic aureole of a granitic stock represent a higher grade attained in the aluminous quartzite. Either the kyanite persists metastably or the kyanite is stable with garnet. If the staurolite-bearing assemblages and the kyanite-bearing assemblages represent the same grade, the phase assemblage is as shown on Figure 5, A2. The location on the petrogenetic grid (Fig. 6, A2) is at a higher grade than A1. A tract has been drawn for these two assemblages (A1, A2) but the position of the arrow is significant only in its relative position to that postulated for other areas (Fig. 6).

Region B – Lower Amphibolite Region

Region B, near Hayes River (Figs. 1, 7), is at lower amphibolite grade. The nature of the contact between the bordering granitoid gneiss and the Prince Albert Group is uncertain. A possible unconformity 15 km southwest of Region B may indicate that the granitoid gneiss underlies the supracrustal rocks. In region B the Prince Albert Group consists of metamorphosed quartzite, pelitic mica schist, iron-formation, and basic magnesian (komatiitic) flows and sills. Small gabbro stocks are emplaced within the volcanic rocks. These units were deformed into a tight syncline-anticline pair and later cut by a gabbro dyke.

Quartzitic rocks contain muscovite, chromian muscovite, andalusite, and accessory minerals. Pelitic rocks are also exposed along this belt; their mineral assemblages are plotted on Figure 5, B1. The assemblage staurolite-biotite-cordierite-garnet-quartz-muscovite ± chlorite, and the same assemblages, but lacking staurolite, are found in the mica schist.

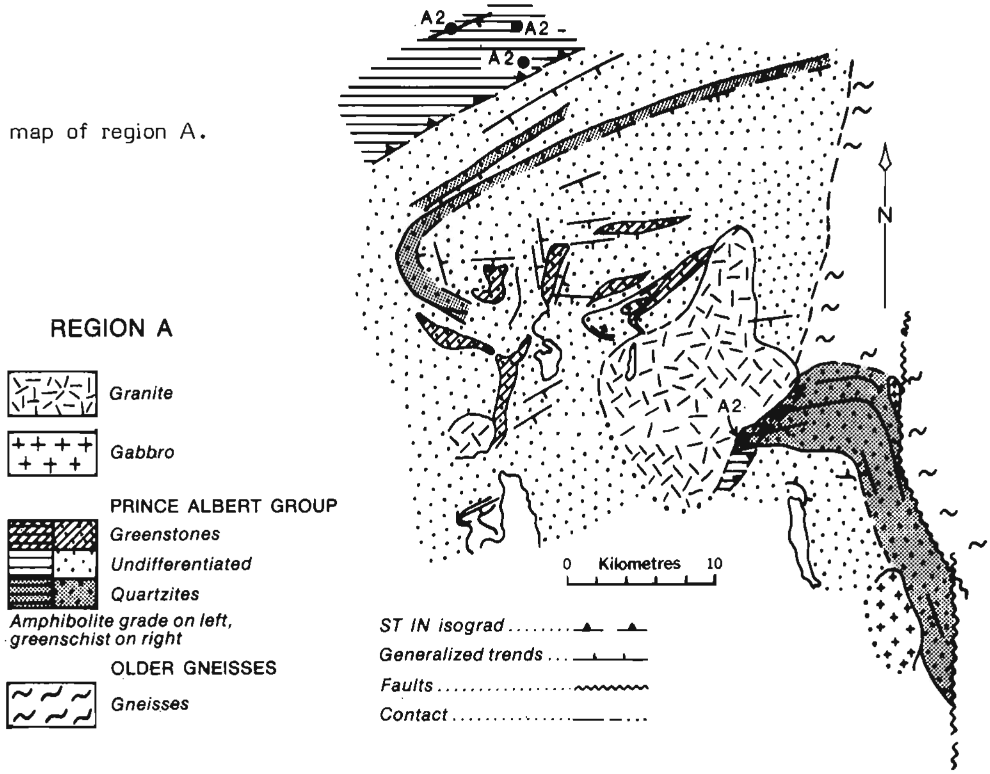
The position of the point defined by the above mineral assemblages is determined by the coexistence of minerals at equilibrium. If the assemblage staurolite-chlorite-cordierite-almandine-muscovite-biotite is stable it can be represented at invariant point B1y (Fig. 6). This is considered to be the most probable case. If, however, the chlorite is a secondary alteration product the assemblage may be represented by points along the univariant line B1x (staurolite + biotite = cordierite + almandine + muscovite). Manganese could have stabilized garnet growth in which case the assemblage would lie along the univariant reaction B1z (chlorite + muscovite = biotite + staurolite + cordierite). However garnets appear in manganese-poor rocks (MnO=0.2 wt %) as well as in manganese-rich rocks (MnO=1.4 wt %).

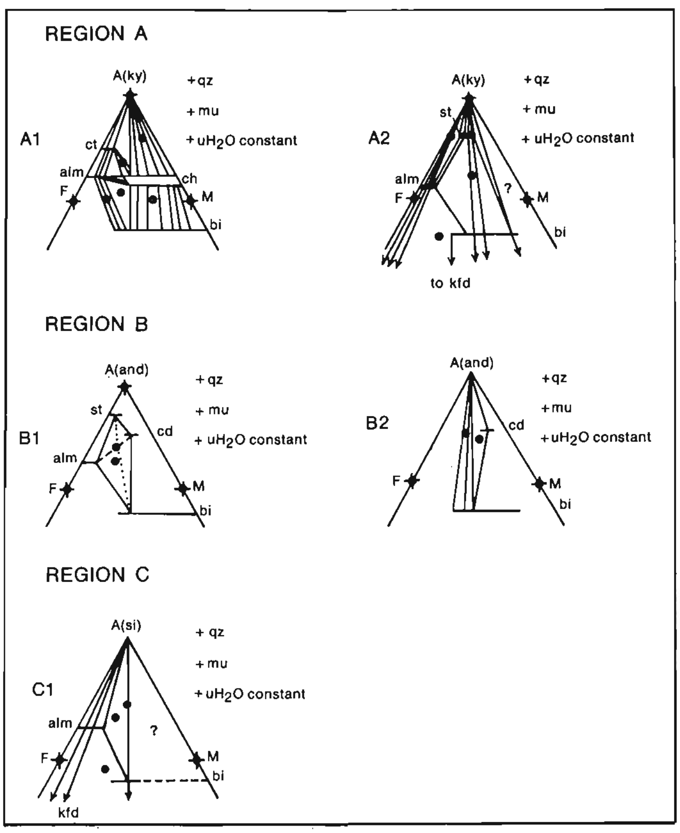
In region B (southwest of the B1 occurrence), the assemblage andalusite-biotite, and rarely cordierite (B2, Fig. 5) represents a higher grade metamorphism than assemblage B1. The arrows representing the three tracks are meaningful only in a relative sense (Fig. 6). Although the temperature ranges may be the same for regions A and B the mineral assemblages of the phyllite of region A record a higher pressure than do those of the mica schist of region B.

Region C – Upper Amphibolite Region

In the upper amphibolite region (Figs. 1, 8) rocks of the Prince Albert Group are conformable with or intruded by granitoid gneiss. The group is composed of mica schist and paragneiss with thick layers of quartzite and thin layers of iron-formation and ultramafic rock (Campbell, 1973; Wolff, 1974) interleaved with pegmatite. The beds have moderate dips and some are overturned.

Figure 4.
Geological sketch map of region A.





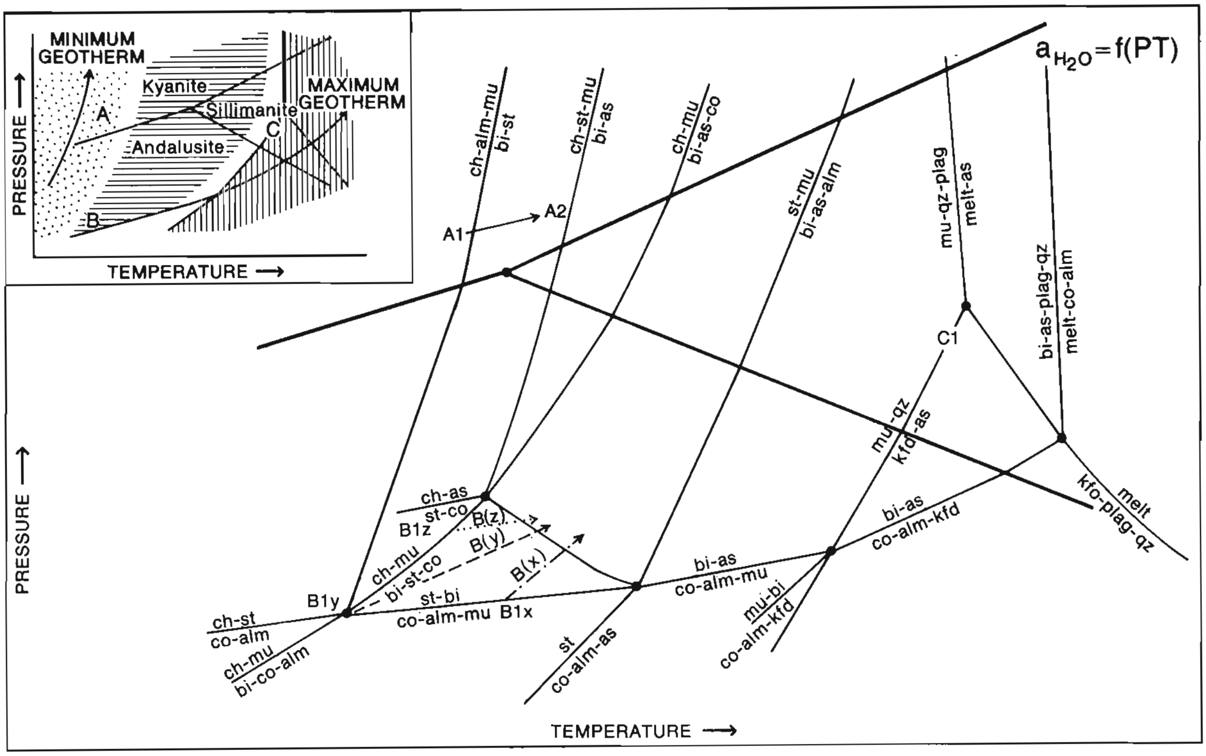
qz	= quartz	alm	= almandine
mu	= muscovite	ch	= chlorite
ky	= kyanite	bi	= biotite
and	= andalusite	cd	= cordierite
si	= sillimanite	st	= staurolite
ct	= chloritoid	kfd	= potash feldspar

Figure 5. AFM plots (Thompson, 1957) for each region. Assemblages are discussed in text.

The quartzite contains muscovite, in part chromian, sillimanite, K feldspar, and accessory minerals including tourmaline. The majority of metamorphosed pelitic rocks consists of quartz-plagioclase-K feldspar-muscovite-biotite-sillimanite gneiss, however quartz-K feldspar-muscovite-biotite-sillimanite-garnet gneiss also occurs. K feldspar, muscovite, and fibrolitic sillimanite commonly occur near each other in thin section (Wolff, 1974). This assemblage represents a breakdown in muscovite and can be represented by any point (C1, Fig. 6) on the "univariant" reaction line muscovite + quartz = sillimanite + K feldspar in the petrogenetic grid (Fig. 6). Chatterjee and Froese (1975) have shown that it is not a univariant line, but rather a "smear" in P-T space spreading over about 30°C because of the solid solution of paragonite in muscovite. There are two kinds of pegmatite in this region, one of which is a muscovite-bearing pegmatite found mainly associated with Prince Albert Group gneiss, and another, which is a pink K feldspar-quartz aggregate, which is found near the gneiss that bounds the Prince Albert Group. Migmatite segregations are not widely developed. Thus, conditions in region C probably lie on the muscovite-quartz reaction very near the field of melting of granitic compositions. The relative position of the point (Fig. 6) outlined by the region C is important in that it indicates higher pressure and temperature than point B and higher temperature than point A.

Grid Calibrations and Tectonic Models

To calibrate a petrogenetic grid and translate the axes into temperature and depth a thorough understanding of the behaviour of silicate minerals under metamorphic conditions is required. The chemical simplicity of the aluminosilicate minerals would suggest that their phase relations should be freer from ambiguity and therefore easier to apply than the relations of more complicated minerals. Unfortunately, this is far from the case. In 1969 an entire issue of the American Journal of Science (vol. 269, p. 257-422) was devoted to the aluminosilicate minerals, and since that time, additional papers on these minerals have been published. Consequently



alm	=	almandine	kfd	=	potash feldspar
as	=	aluminosilicate	mu	=	muscovite
bi	=	biotite	plag	=	plagioclase
ch	=	chlorite	qz	=	quartz
co	=	cordierite	st	=	staurolite

Figure 6. Uncalibrated petrogenetic grid (Ermanovics and Froese, 1978). Regions A, B, and C are plotted on inset showing fields of greenschist (dots), lower amphibolite (horizontal lines), and upper amphibolite (vertical lines). Also shown are maximum and minimum geotherms deduced for Prince Albert Group belt. B(X), B(Y), B(Z), A1, A2, B1, B2, and C1 are explained in text.

the use of phase relations of the aluminosilicates is complicated by the necessity for choosing from among several alternative models (Fig. 9). In a recent compendium reviewing in part the metamorphic petrology of selected groups of rocks, based on the application of experimental data (Bailey and McDonald, 1976), three different triple points of aluminosilicates were proposed. Newton and Fyfe (1976) supported Turner's (1968) scheme which has a triple point at about 2 kb* and 450°C. Greenwood (1976) reviewed the experiments of Richardson et al. (1969) and Holdaway (1971) and concluded that the latter was more convincing. Greenwood (1976) tentatively proposed a new triple point about 50°C higher than Holdaway's triple point which satisfies petrographic observations better than any of the previous proposals. The third author in this compendium (Schreyer, 1976) preferred the triple point of Richardson et al. (1969) because he found that field and paragenetic studies favoured it. The most recent paper available to the author (Anderson et al., 1977) supports Holdaway's triple point.

The difficulty of determining the equilibrium of the aluminosilicate minerals stems from the fact that the free energy of reaction is only a few hundred calories per mole in contrast with dehydration reactions which are tens of kilocalories per mole (Greenwood, 1976). The entropy of reaction is likewise small (Greenwood, 1976). Richardson et al. (1969) and Greenwood (1976) have pointed out that this small entropy change makes equilibrium temperature very sensitive to changes in the free energy of the phases involved. These perturbations can be caused by additional components, grain size, strain energy and aluminosilicate disorder. The effect of small amounts of extra components will depend upon the fractionation of these components between co-existing aluminosilicate polymorphs (Albee and Chodos, 1969; Chinner et al., 1969). In addition to the variability of composition, it is possible that the state of ordering of aluminum and silicon tetrahedral co-ordination in sillimanite may have an equally large effect (Zen, 1969; Holdaway, 1971; Greenwood, 1976).

Triple point models deduced for ideal systems, inferred from the results of controlled experiments and from observations in natural systems are significantly different. Ideally, well characterized aluminosilicate polymorphs should exist at a triple point. Operationally, however, the experimental systems employed have more variables than pressure and temperature and this has led to the establishment in experiments of fields of stability of each of the polymorphs. Different operations have yielded different fields of stability (Richardson et al., 1968, 1969; Althaus, 1969; Newton, 1969). Textural relations in rocks suggest that the type of energy flux and variations in composition and/or structural perfection and other variables as yet unknown may be more important than the relationships between variables recognized by classical thermodynamics, so that the "metastable" formation (Hollister, 1969) and "meta-stable persistence" (Pitcher, 1965) of each of the aluminosilicates may be important in nature. Thus, the equilibrium phase diagram considering P and T as the only variables may inadequately represent laboratory results or field occurrences. Field geologists have usually preferred low

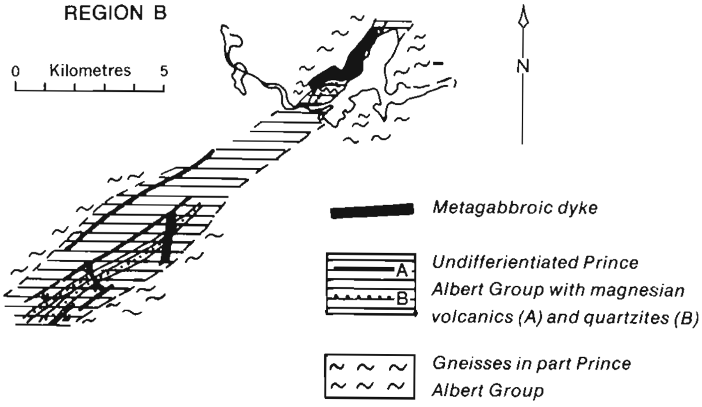


Figure 7. Geological sketch map of region B.

pressure triple points (Rutland, 1965; Pitcher, 1965; Turner, 1968; Verhoogen et al., 1970), but others, using independent pressure and temperatures estimates, postulated higher temperature and pressure triple points (Rosenfeld, 1969; Hietanen, 1969). The conclusion is that no single triple point exists in nature, even though time-scales are 10^8 to 10^{10} times longer in nature than in the laboratory.

Experimental work on other mineral assemblages may help in judging the relative positions of the three mapped regions of the District of Keewatin (Fig. 1). Experimental evidence (Schreyer and Seifert, 1969a,b) shows the assemblage magnesian chlorite-kyanite-quartz to be a high pressure equivalent of cordierite. The stability field of this assemblage can be represented on a P-T diagram on the high pressure side of a line that includes the point at 5 kb and 500°C and the invariant point at 6.5 kb and 650°C. The lower boundary of this assemblage is presumably marked by the upper stability limit of pyrophyllite. Since pyrophyllite forms from clays near 300°C (Hoscheck 1969) the breakdown of pyrophyllite can never be less than this temperature. The upper stability curve of pyrophyllite is currently under discussion (Kerrick, 1968; Althaus, 1966; Haas and Holdaway, 1973). Shown in Figure 9 is the breakdown of pyrophyllite according to Althaus (1966). The usual assumption is that the partial pressure of water equals total pressure. Khlestov (1974a) preferred fluid pressure estimates for greenschist reactions in the neighbourhood of six tenths of the total pressure. In the pile of pelite, siltstone, and sandstone that makes up the Prince Albert Group an estimate of P_{H_2O} much less than P_{total} seems unreasonable. The higher than minimum pressure values shown in Figure 9 are preferred for the lower boundary of the kyanite field.

The presence of abundant iron in various associated rock types suggests that the lower boundary of the chlorite-kyanite-quartz field may be lower than that reported by Schreyer and Seifert (1969 a,b) because of the enlarging effect of iron on the stability field of chlorite. This would shift all mineral stability fields into lower pressure regions, but the relative positions of mineral assemblages throughout the region would not be affected. If the kyanite near the contact of the small pluton in region A is stable then the contact aureole was formed under relatively high pressures. Holdaway (1971) for instance, would estimate it to be in excess of 6 kb, whereas Atherton et al. (1975) would suggest pressures in excess of 3 kb. In spite of the uncertainties, the pressures attained in region A were probably in excess of 4 to 5 kb and the temperatures were generally about 450°C ranging to a high of 650°C near the pluton.

The invariant point, which may be represented by the mineral assemblages displayed in B1 (Fig. 5), has been calculated by Hess (1969) at 2.7 kb (P_{H_2O}) and 550°C. Thompson (1976) calculated this same invariant point to be about 3.2 kb (P_{H_2O}) and 580°C. In both these studies $\alpha H_2O = 1$.

A garnet- and muscovite-bearing, very coarse grained granite and associated pegmatite, containing graphic granite textures is present about 15 km to the north of region B. The granite contains large inclusions of country rock and is characteristic of epizonal to mesozonal plutons. The graphic granite, the large muscovite crystals, and the abundant tourmaline crystals at the granite contact all suggest that the volatile content of the magma was very high. The coarse grained textures and minerals suggest that it crystallized in an environment where P_{H_2O} was equal to P_{total}. The lowest estimate of pressure and temperature at which water-rich muscovite-granite can crystallize is about 3.3 kb and 650°C. This is estimated from the intersection of the minimum melting of granite and breakdown of muscovite and quartz (Merrill et al., 1970; Carmichael et al., 1974; Chatterjee and Froese, 1975). The most abundant aluminum silicate in the nearby pelite to the south is andalusite, but sillimanite was seen in a locality northwest of the stock where gneiss assumed to be associated with the stock is present. In region B therefore the estimated pressure was approximately 3 kb and the temperature ranged from about 550° to 650°C near plutons.

In region C the temperature, and pressures (fluid and total) are not known well enough to be quantified. In a field study of sillimanite-garnet-biotite-K feldspar-quartz-muscovite assemblages Froese (1973) concluded that the P_{H_2O} was not equal to the total pressure. One possible reason for the lowering of P_{H_2O} is that the water was taken up to produce silicate melt. Thus, the highest grade metamorphic rocks may not reflect the peak conditions of metamorphism, but rather a set of quench conditions. The relative position of region C, as already indicated, is well known.

Although the relative positions of the three regions on a P-T diagram is supported by experimental work, the exact location of each region on a P-T grid is not known. To avoid errors occasioned by ignoring the variability in experimental and natural systems (Zen, 1977), the following procedure, similar to that used by Thompson (1977), was observed.

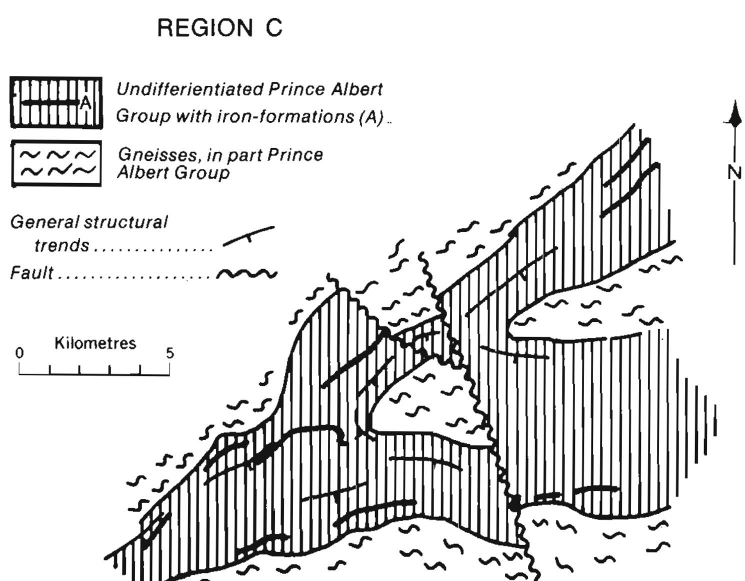


Figure 8. Geological sketch map of region C (F.H.A. Campbell, pers. comm.).

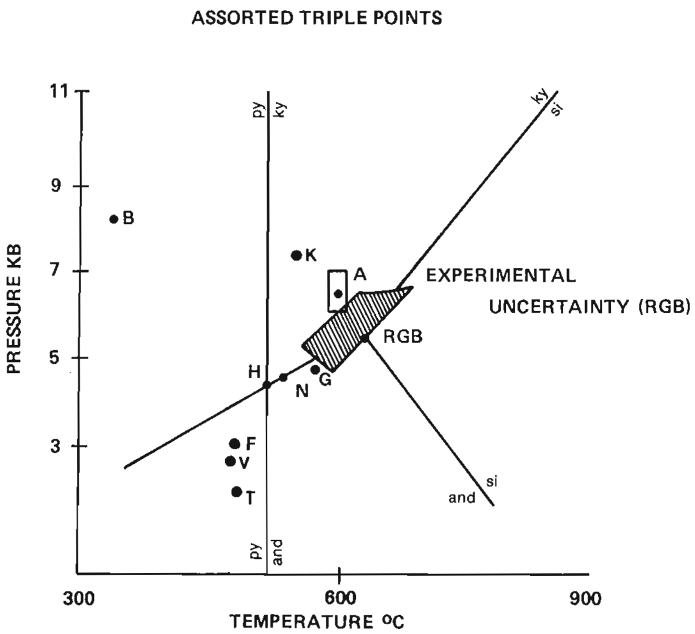


Figure 9. An assortment of triple points. Points labelled according to investigators: (K) (Khlestov, 1974), (A) (Althaus, 1969), (RGB) (Richardson et al., 1969), (N) (Newton, 1969), (H) (Holdaway, 1971), (F) (Brown and Fyfe (1971), (T) (Turner, 1968), (G) (Greenwood, 1976), (V) (Verhoogen et al., 1970), (B) (Bell, 1963). Pyrophyllite (py) breakdown to kyanite (ky) or andalusite (and) after Althaus (1966). (Sillimanite - si).

Two tectonic models representative of widely different pressure-temperature estimates of the aluminosilicate triple point current in the geological literature were developed. One featured a near-surface triple point (Verhoogen et al., 1970; Turner, 1968; Brown and Fyfe, 1971) and the other, a deep triple point (Richardson et al., 1969). The petrogenetic grid employed retains the relative positions of reactions indicated by Ermanovics and Froese (1978) and Hess (1969). The position of the triple point was moved up and down on this grid to conform with each of the two models chosen. The depth and temperature for each of the three regions where the Prince Albert Group was studied were thus estimated and the three points obtained were entered on a distance-depth diagram with erosion traces and contoured isotherms. It should be noted that the various temperature and pressure estimates of critical reactions from the grid are not particularly accurate.

In the deeper model (Fig. 10) the triple point of Richardson et al. (1969) was used. The presence of a thermal high in the northeast is clearly shown. An immense amount of material, especially in the west, has been removed by erosion. Estimates of this thickness range from eight times the stratigraphic thickness of the Prince Albert Group in the west to five times the stratigraphic thickness in the east. Complexly and isoclinally folded Prince Albert Group and old gneiss probably have contributed to this now eroded volume. There is a possibility that low temperature allochthonous material may have contributed to the larger thickness that has been removed from the west. Although the Prince Albert Group is continuous from west to east, a north-south fault system associated with gabbro stocks emplaced in locally ultramafic-rich Prince Albert Group lies between region A and region B. The deduced movement (west side down) on this fault is not great. The nature of this discontinuity between the low grade and the medium grade metamorphic terranes must be clarified through further work.

The type of metamorphism noted in the central and eastern regions is very similar to that described by Thompson (1978) for the Slave Structural Province. In both areas similar types of tectonic models are applied to rocks of about the same age using similar triple points. Thompson interpreted the thermal highs in the Slave Province as having a spacing between 20 and 200 km whereas the thermal highs in this part of the northern Churchill Province were spaced much farther apart. The presence of medium pressure low grade rocks far from thermal domes as observed in region A has no counterpart in the Slave Province, but is a feature predicted for this type of model by Richardson (1970) and has been noted in other Archean regions (Saggerson and Turner, 1976). The erosion surface curve shown in the model diagrams is unusual in that it must include three points that are not co-linear; most erosion surfaces can be represented by straight lines or gently curved P-T loci (Thompson, 1977). In Rhodesia, for example Saggerson and Turner (1976) used the triple point of Richardson et al. (1969) in a diagram similar to Model A (Fig. 10). They suggested that pressure did not vary but some rather high pressure phase assemblages that occur in Rhodesia were not integrated into their model (cf. Chinner and Sweatman, 1968).

A shallow model (Fig. 11) based on the alternate calibration of Verhoogen et al. (1970), yields a much greate thermal gradient than that of the previous model. In deeper model the minimum thermal gradient is 30°C/km, and the maximum 45°C/km; whereas in shallow model the minimum is 58°C/km, and the maximum 80°C/km, at the triple point. A similarly shaped pressure-distance plot results in shallow model although the eroded volume in that model is much less.

The choice between the two calibrations and models is a difficult one. Deeper model adherents will infer extensive allochthony of units and thick crust during the Archean with variable geothermal gradients such as those measured today, whereas shallow model supporters will be compelled to accept thin mobile Archean crust with high geothermal gradients. Read and Watson (1976) have concluded in their general review of the early history of the earth that, although some authors have reasoned that the character of the Archean complexes indicates the dominance of steep geothermal gradients in the Archean crust, it is doubtful whether any general progression from steep to less steep geothermal gradients can be substantiated. The occurrence of kyanite in the Scourian (Read and Watson, 1976), Naqssugtoqidian

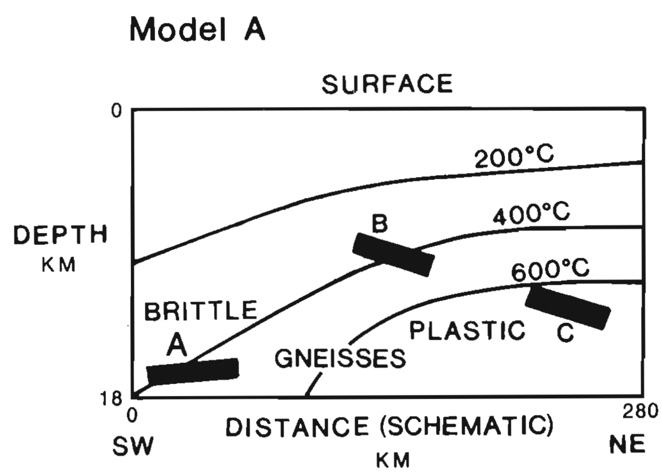


Figure 10. Deeper model – Distance-depth diagram with estimates of temperature added as contours, based on triple point of aluminosilicates by Richardson et al. (1969). Solid lines represent regions A, B, C, as labelled.

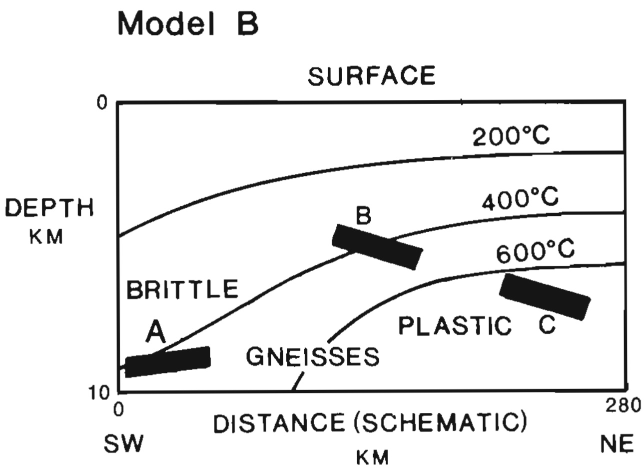


Figure 11. Shallow model – Distance-depth diagram with estimates of temperature added as contours, based on triple point of Verhoogen et al. (1970). Solid lines represent regions A, B, or C, as labelled. Note difference in depth scale with respect to Figure 10.

kyanite in Greenland (Bridgwater et al., 1976), Archean kyanite in Russia (Khlestov, 1974 a,b) and Australia (Joplin, 1968), kyanite schist in Africa (Saggerson and Turner, 1976), the Archean kyanite localities reported from each of the Slave (Thompson, 1978), Churchill (Lewry et al., 1978) and Superior Structural Province (Jolly, 1978) leave little doubt that metamorphic facies were being formed at intermediate pressure during Archean time, no matter which of the above models is applied. The presence of garnet-clinopyroxene-orthopyroxene-plagioclase granulite of probable Archean age 200 km south of the area mapped (Reinhardt and Chandler, 1973; Schau and Hulbert, 1977) also supports the existence of a locally thick crust in the Churchill Structural Province during late Archean time.

CONCLUSIONS

(1) A large Archean thermal high has metamorphosed the Prince Albert Group and the underlying gneiss in the northern District of Keewatin. The "peak or quench" metamorphism outlasted the deformation, as is shown by isograds cutting across regional structures.

(2) Three regions, representative of each of the kyanite-, andalusite-, and sillimanite-bearing portions of the Prince Albert Group were metamorphosed at different pressures.

(3) Later, local metamorphism around small plutons was followed by widespread incomplete greenschist diaphthoresis, which is responsible for the Hudsonian radiometric dates obtained from platy minerals of the region. This metamorphic downgrading may be associated with a phase of the widespread east-northeast fault systems which affected the Churchill Province in post-Archean time.

(4) The lack of agreement on the location of the triple point of aluminosilicates in P-T space results in very different tectonic models derived from metamorphic considerations based on different petrogenetic assumptions. Two such models have been considered. Moderate geothermal gradients and extensive deep erosion and a thick Archean crust are indicated by one; high geothermal gradients, but much less erosion and a thin Archean crust, are indicated by the other. To choose between them in the District of Keewatin is not possible on the basis of work done to date. The presence of differentiated sediments argues for current rates of chemical weathering on a stable craton (deeper model) or very rapid rates of chemical weathering on an unstable crust (shallow model). Metamorphic indications of deep crustal metamorphism about 200 km south of the mapped area favour a deeper model. The subsequent history of deformation, metamorphism, cooling and erosion, widespread faulting, dyke emplacement, local plutonism, and eventual uncovering to mark the end of the Hudsonian event, yield no definitive data to support either model.

The eventual choice of one model over the other will result not only in better understanding of the complex metamorphic systems here discussed, but also in the solution of the detailed structural history of the area. Currently, I favour a deeper model rather than a shallow model, mainly because of its uniformitarian simplicity.

ACKNOWLEDGMENTS

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METAMORPHISM OF THE LABRADOR TROUGH

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Dimroth, E. and Dressler, B., Metamorphism of the Labrador Trough; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 215-236, 1978.

Abstract

The Labrador Trough, an erosional remnant of an Early Proterozoic (Aphebian) geosyncline in northeastern Quebec and Labrador, was folded and metamorphosed during the Hudsonian Orogeny. Metamorphic grade increases from west to east across the trough, and isograds are subparallel to the sedimentary and tectonic strike. The sequence of metamorphic isograds suggests that metamorphism took place at approximately normal geothermal gradients. Domes in the basement were subjected to high grade metamorphism during the Kenoran Orogeny, and were subjected to retrograde metamorphism during the Hudsonian Orogeny.

Very low grade metamorphism is pre- and synkinematic, low grade metamorphism is synkinematic and medium to high grade metamorphism is late- to postkinematic. Tectonic relations and K-Ar age determinations suggest that the late- to postkinematic high grade metamorphism in the eastern part of the trough took place earlier than the synkinematic low grade metamorphism in the west. Metamorphism in the Labrador Trough is possibly due to tectonic subsidence followed by complete thermal readjustment and uplift with no abnormal heat flow.

Résumé

La fosse du Labrador, un géosynclinal profondément érodé du Protérozoïque inférieur (Aphébien) située dans le nord-est du Québec et au Labrador, a été plissée et métamorphisée au cours de l'orogénèse hudsonienne. Le degré de métamorphisme augmente de l'ouest à l'est à travers la fosse et les isogrades sont subparallèles aux limites des faciès sédimentaires et des zones tectoniques. La succession des isogrades métamorphiques suggère que le métamorphisme s'est produit à des gradients géothermiques à peu près normaux. Des dômes du socle métamorphisés à un faciès élevé au cours de l'orogénèse kénoranienne ont été ensuite soumis à un métamorphisme rétrograde au cours de l'orogénèse hudsonienne.

Le métamorphisme est pré- et syn-cinématique où il est faible, tandis qu'il est tardi- et post-cinématique où il est fort. Les relations tectoniques et les âges au K-Ar suggèrent que le métamorphisme élevé, tardi- à post-cinématique, à l'est de la fosse s'est produit avant le métamorphisme faible pré- à syn-cinématique de la partie ouest de la fosse. Le métamorphisme de la fosse du Labrador est possiblement dû à la subsidence tectonique suivi du réajustement thermique complet et d'un soulèvement, sans flux thermique anormal.

INTRODUCTION

The Labrador Trough, a north-northwest-trending belt of Aphebian rocks in northeastern Quebec and Labrador (Fig. 1) separates the Superior Province to the west from the Churchill Province to the east; it is probably the erosional remnant of a formerly more extensive geosyncline (Dimroth et al., 1970). The Aphebian geosynclinal rocks overlie a basement of Archean gneiss, migmatite and granite. Archean gneiss and migmatite were metamorphosed to upper amphibolite and granulite facies during the Kenoran orogeny, about 2500 Ma ago. They were intruded by diabase dykes 2200 Ma ago (Fahrig and Wanless, 1963).

The geosynclinal filling of the Labrador Trough probably is of late Aphebian age, younger than the diabase dykes but older than 1850 Ma (Fryer, 1972; Dressler, 1975). It is markedly asymmetric. In the west, the Aphebian sequence is less than 3 km thick; sedimentary rocks predominate, and comprise a high proportion of mature deposits (orthoquartzites, dolomites, iron-formation). In the centre, the sequence is more than 6 km thick and is mainly composed of mafic igneous rocks. In the extreme east the sequence appears to be fairly thick, and sedimentary rocks again predominate, but are less mature than the sediments in the west; furthermore, some mafic igneous material is present.

The Aphebian rocks of the Labrador Trough, and parts of their basement, have been folded, faulted, and metamorphosed during the Hudsonian Orogeny, about 1900-1600 Ma ago (Wanless et al., 1968). Overfolding and thrusting are directed to the west but the intensity of the deformation and the grade of metamorphism increased eastward across the fold belt. Thus, the Labrador Trough shows the sedimentary, tectonic, and metamorphic gradients characteristic of Alpine-type geosynclines (Aubouin, 1965).

Fold axes generally plunge southward, hence the deepest level of the geosyncline is exposed in the north, and the shallowest level in the south. It is therefore possible to study the nature of the Superior-Churchill boundary at various levels within the Proterozoic sequence and in the Archean basement. A few late Hudsonian granite stocks and pegmatite dykes occur on the extreme east side of the Labrador Trough.

The geosyncline was deeply eroded after the Hudsonian Orogeny. The deposition of the Sims Formation and the intrusion of the Shabogamo Gabbro occurred between 1100 and 1600 Ma. Rocks of the southernmost Labrador Trough, and the overlying Sims Formation and Shabogamo Gabbro were involved in renewed folding, faulting, and metamorphism during the Grenville Orogeny, between 1100 and 800 Ma (Wynne-Edwards, 1960, 1961; Fahrig, 1967).

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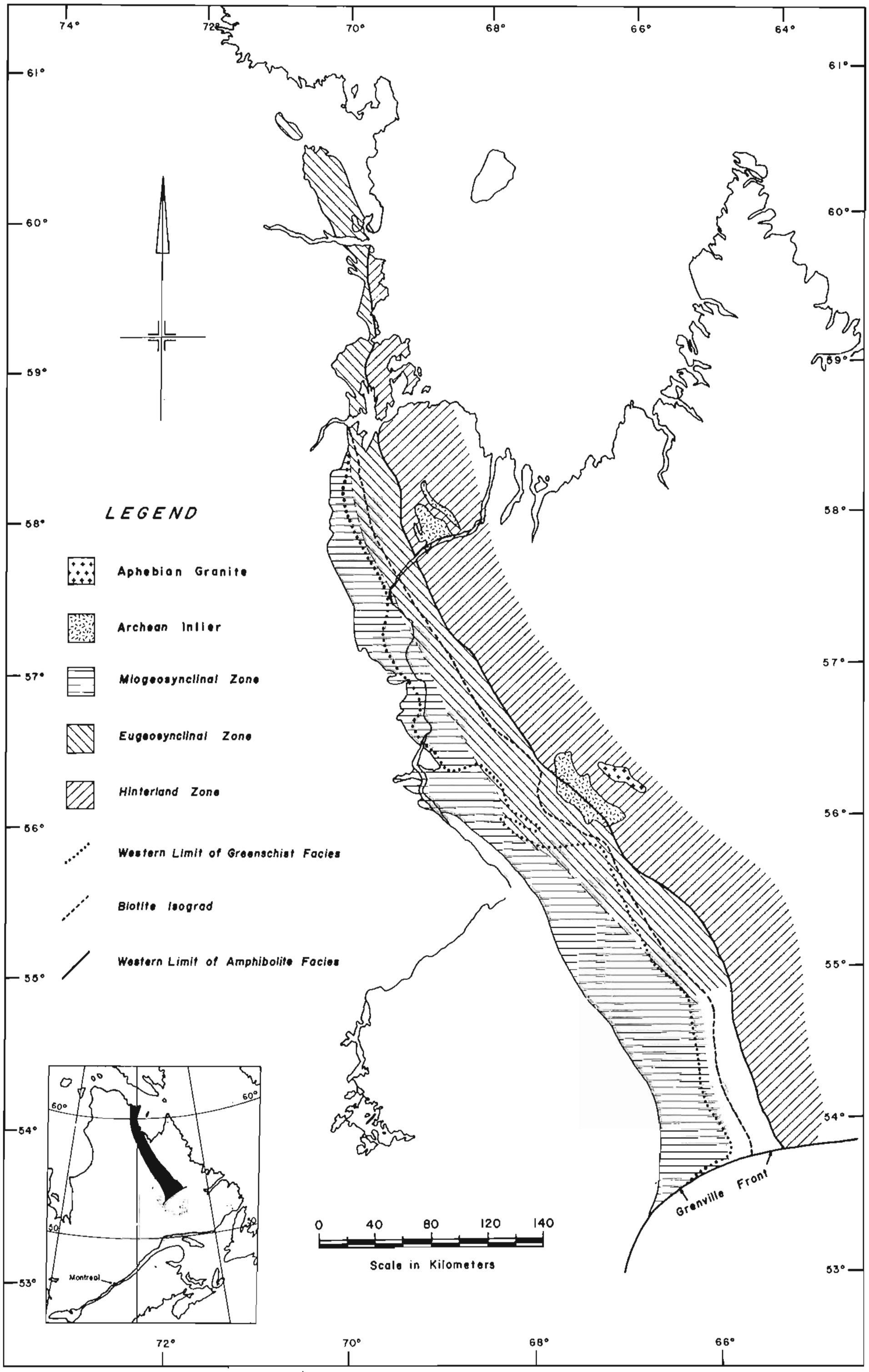


Figure 1. Location and generalized metamorphic map of the Labrador Trough.

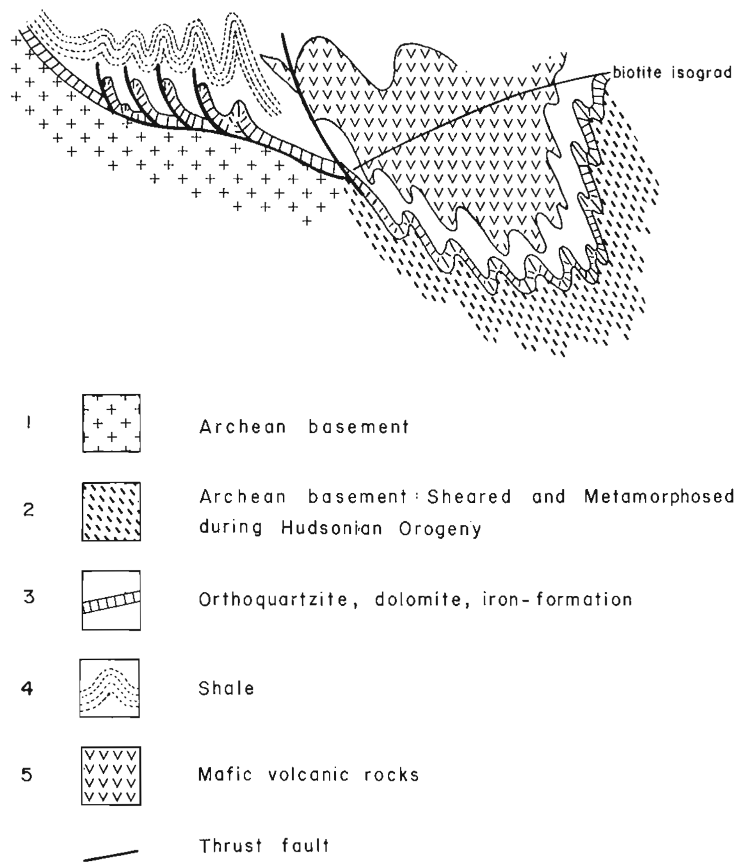


Figure 2. Tectonic relations at the boundary between the Churchill and Superior provinces. Note change of expression of boundary with level of exposure. Lithologic units numbered in order of decreasing age: 1-2, Archean; 3-5, Aphebian.

Basis of Compilation

This compilation is based on reconnaissance mapping, therefore great accuracy should not be demanded of it. Isograds were mapped in moderate detail in the central Labrador Trough (Baragar, 1967; and this paper) and in the Fort Chimo region (Sauvé et Bergeron, 1965; Gélinas, 1965). Elsewhere isograds must be approximated from rock descriptions.

Tracing of isograds is hampered by insufficient mapping, lack of petrographic work, and by unsuitable lithologies. The boundary between the subgreenschist and greenschist facies is approximate north of 56°30'N because mafic igneous rocks are absent; tracing of the boundary between the greenschist and amphibolite facies is possible only where numerous determinations of the anorthite content of plagioclase are available; isograds within the amphibolite facies can rarely be delineated with precision because aluminous pelites are uncommon in most areas.

The important question of which minerals are in stable co-existence and which are not has rarely been discussed in the literature of the region. Our information on subgreenschist facies parageneses is largely drawn from Baragar (1967) and Ott (1972). Information on very low grade parageneses in iron-formations is in part conflicting (Dimroth and Chauvel, 1973; Zajac, 1973; Klein and Fink, 1976). Greenschist facies assemblages were noted by Baragar (1967), Bérard (1965), Sauvé et Bergeron (1965), and Hardy (1968). Metamorphic relations in the amphibolite and granulite facies have been the subject of a major monograph (Gélinas, 1965).

This paper is concerned with the Hudsonian metamorphism of that part of the Labrador Trough within the Churchill Province. Dressler is responsible for the compilation and description of the metamorphism between 56°30'N and 57°30'N; the senior author compiled the remainder, and wrote the paper.

Definition of Province Boundaries

The westernmost limit of intense Hudsonian deformation is here defined as the boundary of the Churchill and Superior provinces, or as the "Hudsonian Front". The physical expression of the Hudsonian Front (or front zone) varies with the level of exposure and lithology (Fig. 2). At the highest exposed level, in competent lithologies (quartzite, dolomite, iron-formation), it is a zone of closely spaced thrust faults dipping steeply east and of narrow folds overturned to the west and, in thick shale sequences, a zone of gradually increasing fold intensity. At somewhat deeper levels, a flat thrust fault within the Aphebian sequence or a décollement at its base defines the Hudsonian Front.

North of Leaf Bay, at the deepest exposed level, the Archean basement has been involved in the Hudsonian deformation. There the Hudsonian Front is situated within the Archean gneissic terrane west of the Labrador Trough. The nature of this front will be discussed later.

The Grenville Front has been drawn at the northern limit of northeast-trending structures recognizable on present maps. It is a zone of northeast-trending thrust faults, shear zones, and folds, superposed upon the north-northwest trending folds of the Labrador Trough, and affecting the Helikian Sims Formation and Shabogamo Gabbro (Wynne-Edwards, 1960, 1961; Fahrig, 1967).

METAMORPHISM OF APHEBIAN ROCKS

Pre-Hudsonian Metamorphism of Aphebian Rocks

The Aphebian rocks of the Labrador Trough were partly metamorphosed before the onset of the Hudsonian Orogeny. There is some evidence that sediments and mafic igneous rocks underwent a static load metamorphism in subgreenschist facies; local spilitization of basalts and gabbros may be related to this pre-orogenic phase of metamorphism. In addition, rocks below basalt flows and in contact with gabbro sills have been locally thermally metamorphosed.

Load Metamorphism

Iron-formation, unconformably overlying the Archean basement and not involved in Hudsonian thrusting and folding, contains unoriented minnesotaite and stilpnomelane; both minerals have been reoriented where a schistosity is present. This suggests that both minerals were already stable when the orogeny began.

Basalts here and there contain amygdules lined with pumpellyite, which may indicate that pumpellyite formed during amygdule filling. Locally, tholeiitic basalts and gabbros have been converted to soda, and in certain cases, potash-rich spilites (Ott, 1972; Dimroth, 1971) during very low grade metamorphism. Spilitization appears to be independent of deformation, and appears to have affected particularly rocks that were permeable during diagenesis (=hyaloclastites) or rocks that are set in voluminous sedimentary material as for example the gabbros at Magnetite (56°30'N, 68°47'W) and Minowean Lakes (56°28'N, 68°30'W). These relations suggest that spilitization took place before the orogeny.

Table 1

Evidence for thermal metamorphism in contact with gabbro sills

	Minerals and texture present	Presumed precursor
	1) Spots 5-20 mm across sub-spherical or lath-shaped, composed of sericite or sericite + chlorite.	Andalusite and/or cordierite.
	2) Spots, several millimetres across composed of albite, in a very albite rich sericite schist.	Spotted adinole.
Marble	Tremolite exhibiting thick prismatic habit, crystal angles and cleavages of pyroxene.	Diopside marble.
Calc-silicate rock	1) Tremolite-talc-calcite schist.	Calc-silicate rock.
	2) Reaction rims of tremolite at margin of carbonate pebbles in conglomerate.	Calc-silicate rock derived from carbonate conglomerate.

Contact Metamorphism

The voluminous pre-orogenic gabbro intrusions and basalt flows caused minor contact thermal metamorphism. Intense retrograde metamorphism of the resulting hornfelses occurred during the Hudsonian regional metamorphism (Table 1).

Most pelites in contact with gabbro sills or underlying basalt flows show no visible trace of contact metamorphism. Perhaps they were still largely water logged at the time of intrusion and, therefore, did not attain a high temperature. Spotted schist is present here and there; the spots are several millimetres to about 2 cm across, and are now composed of sericite and/or chlorite. Shapes of the spots are not suggestive of any particular mineral but they may be derived from andalusite and/or cordierite. Soda-rich quartz-albite-sericite-chlorite schist is derived from adinoles.

Diopside developed in marbles at the contacts between dolomite and gabbro was converted to tremolite during the regional metamorphism, with preservation of crystal shapes and pyroxene cleavages. Calcite-tremolite-talc schist is derived from calc-silicate rocks that formed during contact metamorphism. Reaction zones formed in conglomerate which contain dolomite pebbles in a silicate matrix at, for example, Ribero Lake (56°50'N, 67°50'W) (Baragar, 1967, p. 21).

Table 2

Typical minerals and mineral associations of the subgreenschist facies

Mafic rocks	Magmatic minerals: Clinopyroxene, plagioclase, rare brown hornblende. Metamorphic minerals: Pumpellyite, prehnite, chlorite, albite, epidote; pumpellyite, chlorite, actinolite, albite, epidote; in Fe-rich gabbros stilpnomelane. Texture: mainly magmatic, with minor metamorphic overprint (see Table 4).
Iron-formations	Diagenetic minerals: Calcite, ankerite, siderite, greenalite, magnetite, hematite, chert. Metamorphic minerals: Minnesotaite, stilpnomelane, chlorite, riebeckite, talc. Associations: Hematite - magnetite ± ankerite ± calcite ± iron silicate or Magnetite ± siderite ± ankerite ± calcite ± iron silicate. Texture: Sedimentary with strong "diagenetic" overprint.
Calcareous rocks	Diagenetic minerals. Texture: Strong "diagenetic" recrystallization.
Other sedimentary rocks	Terrigenous detrital components: Quartz, feldspar. Metamorphic minerals: Sericite, chlorite. Texture: Generally sedimentary with diagenetic overprint. Some schistosity in pelite; sericitization of feldspar of arkoses in shear zones, deformation textures in some silica-cemented orthoquartzite.

Hudsonian Metamorphism of Aphebian Rocks

Subgreenschist Facies

Typical minerals and mineral associations of the subgreenschist facies are listed in Table 2. Generally, a mixture of "primary" (magmatic and diagenetic) and metamorphic minerals is present. Prehnite and/or pumpellyite occur in mafic rocks and define the facies. Lawsonite and zeolites are absent.

The subgreenschist facies may be subdivided into two zones defined by the associations pumpellyite-prehnite-chlorite and pumpellyite-actinolite. Ott (1972) described spilites containing pumpellyite, prehnite, and chlorite from Minowean Lake. A little farther east, pumpellyite is accompanied by some actinolite (and chlorite), but prehnite is absent. The reaction isograd

prehnite + chlorite = pumpellyite + actinolite + quartz

has not been mapped.

The "primary" (magmatic and diagenetic) textures of the rocks have been somewhat overprinted by new mineral growth in subgreenschist facies rocks. The successive

Table 3

Typical minerals and mineral associations of the greenschist facies

	Chlorite-zone	Biotite-zone	Garnet-zone
Pelitic rocks	Quartz - albite - muscovite - chlorite ± epidote ± actinolite. Texture: Sedimentary texture preserved in lower grade; phyllite in higher grade.	Quartz - albite - muscovite ± biotite ± epidote ± actinolite. Texture: Phyllite and mica-schist.	Quartz - albite - muscovite ± biotite ± garnet ± epidote ± actinolite.
Iron-formation	Minnesotaite, stilpnomelane, riebeckite, talc, siderite, ankerite, calcite, magnetite, hematite.	Grunerite, stilpnomelane, ankerite, siderite, calcite, hematite, magnetite, locally riebeckite, acmite, actinolite.	Grunerite, garnet, ankerite, siderite, calcite, magnetite, hematite.
	Texture: Recrystallized		
Impure dolomites, marls	Talc, tremolite.	Phlogopite, talc, tremolite.	Phlogopite, talc, tremolite.
	± calcite ± dolomite ± quartz ± albite.		
Mafic rocks	Albite - epidote - actinolite - chlorite ± sphene, stilpnomelane in Fe-rich rocks.		
	Texture: Magmatic texture with minor metamorphic overprint.	Metamorphic texture	
Ultramafic rocks		Magmatic: Olivine, clinopyroxene. Metamorphic: Serpentine, talc, actinolite, magnetite. Texture: Magmatic texture with minor metamorphic overprint.	

overprinting of primary textures are discussed in some detail later. Subgreenschist facies rocks underlie a strip in the west of the Labrador Trough from the Grenville Front to north of 56°30'N. It is difficult to trace the upper boundary of the subgreenschist facies farther north, because the critical lithology — mafic igneous rock — is absent. It should be noted, however, that the sharp westward bend in the boundary of the subgreenschist facies at 56°30'N probably is nonexistent and simply reflects the absence of critical lithologies north of this latitude.

Greenschist Facies

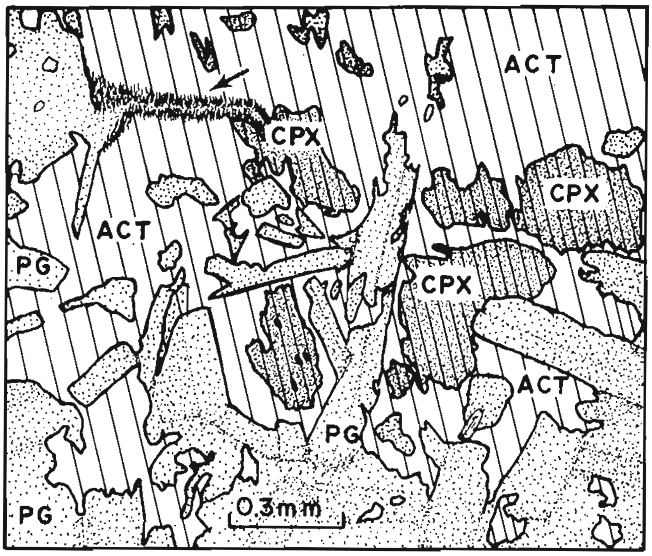
Mineral associations of the greenschist facies are listed in Table 3. The upper limit of the subgreenschist facies is defined by the disappearance of pumpellyite in mafic igneous rocks. Most of the magmatic clinopyroxene is replaced by both actinolite and chlorite. This hydration reaction is strongly dependent on local water vapour pressure and is discussed later. The upper limit of the greenschist facies is defined by the appearance of plagioclase (oligoclase) in mafic igneous rocks.

The biotite and garnet isograds subdividing the greenschist facies were both defined in pelitic sedimentary rocks. In mafic rocks garnet first occurs in the amphibolite facies. Amphibole of mafic igneous rocks is an actinolitic hornblende (Table 4). It is also evident that some reaction occurred between albite and the actinolitic hornblende, since amphibole contains blue-green pleochroic rims where both minerals are in contact.

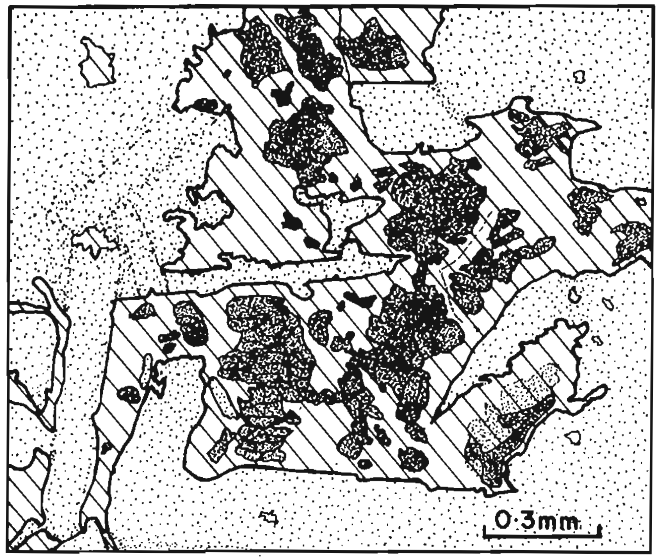
Textures in low grade metamorphic rocks

Textural overprinting is a function of grain growth, deformation, and the intensity of hydration reactions. At very low temperatures, nucleation of new minerals is induced, leading to growth of metamorphic minerals as replacement of "primary" minerals and as overgrowth of "primary" minerals. At higher temperatures, nucleation of new minerals is more abundant and leads to a crystalloblastic fabric. Deformation leads to granulation of earlier crystals, to the parallel orientation of minerals, and to the formation of schistosity. Hydration reactions transform pyroxene to amphibole and feldspar to sericite. These interacting effects, as observed in the textural overprint of gabbros and the effects of sericitization of feldspar upon the earlier textures, are discussed below. General effects of textural overprint are summarized in Table 5.

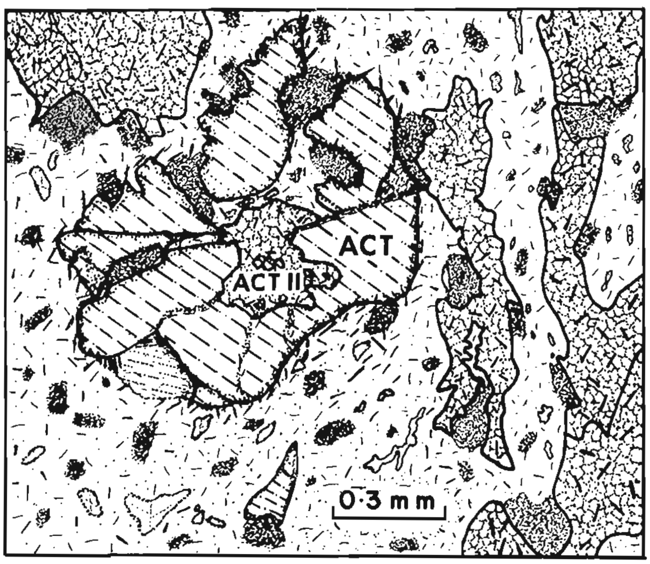
Ophitic gabbroic rocks originally consisted of clinopyroxene, plagioclase laths, and minor orthopyroxene and olivine. In the pumpellyite-prehnite facies, orthopyroxene and most olivine was replaced by chlorite and/or serpentine whereas most of the clinopyroxene is preserved. Oriented replacement (b- and c-axes parallel) of clinopyroxene by actinolite (amphibole I) first occurs in the uppermost subgreenschist facies. Simultaneously, fibrous amphibole (amphibole III) nucleated at the surface of the large actinolite grains as an overgrowth and pierced the plagioclase laths. Finally, relicts of clinopyroxene enclosed in amphibole I have been replaced by fine prismatic actinolite (amphibole II) or by chlorite. Plagioclase has been replaced by albite dusted with clinozoisite and corroded by chlorite. The sequence of textures is illustrated in Figure 3.



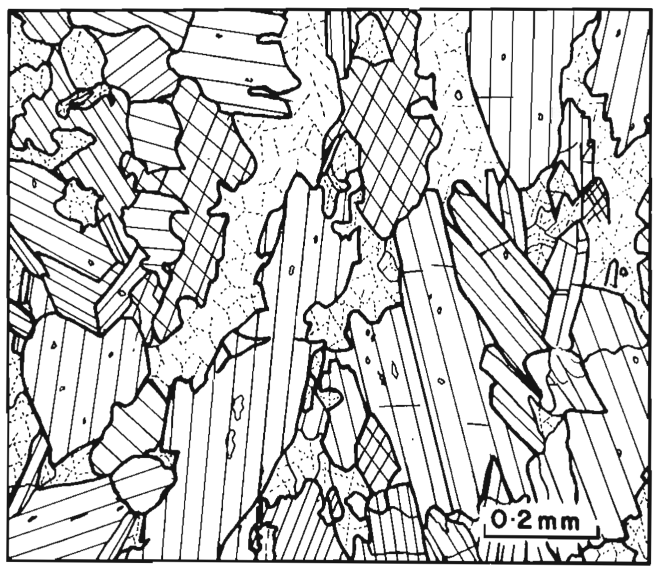
A) Lowermost chlorite zone: Partial replacement of clinopyroxene (CPX) by actinolitic hornblende (ACT). Saussuritization of plagioclase (PG). Some overgrowth of fibrous actinolite on actinolitic hornblende (arrow). Section B18-5, Romanet Lake.



B) Chlorite zone: Remnants of clinopyroxene have been replaced by fine grained prismatic actinolite (densely stippled). Plagioclase and actinolitic hornblende patterned as in 3(A). Section B9-6, Romanet Lake area.



C) Transition chlorite zone-biotite zone: Actinolitic hornblende (ACT) after clinopyroxene slightly deformed and recrystallized. Remnants of the fine grained prismatic actinolite (ACT II) replacing remnants of clinopyroxene within amphibole I are preserved. Saussuritized feldspar has recrystallized to well defined crystals of clinozoisite (densely stippled) and albite (clear). Section A31-12, Romanet Lake area.



D) Amphibolite facies: Recrystallization to schistose amphibolite. Section B29-3, Romanet Lake area.

Figure 3. Successive stages of metamorphism of metagabbro.

Table 4

Composition and optical properties of amphibole in metagabbro (chlorite zone) of the central Labrador Trough

				CATIONS ON BASIS OF 23 O-ATOMS			
	B4-5	A35-7		B4-5		A35-7	
SiO_2	54.87	48.01	Si	7.89	8.00	7.29	8.00
Al_2O_3	1.71	6.46	Al^4	0.11		0.71	
Fe_2O_3	4.20	2.76					
TiO_2	0.48	0.61	Al^6	0.18		0.42	
			Fe^{III}	0.45	0.68	0.31	0.80
FeO	11.42	16.58	Ti	0.05		0.07	
MgO	12.55	0.85					
CaO	11.44	10.34	Fe^{II}	1.37	4.08	2.10	4.34
Na_2O	0.58	0.85	Mg	2.71		2.25	
K_2O	0.10	0.13	Ca	1.76		1.67	
H_2O	2.15	3.31	Na	0.16	1.94	0.22	1.91
TOTAL	99.50	98.90	K	0.02		0.02	

	B4-5	A35-7		B4-5	A35-7
n_x	1.633	1.641			
n_y	1.644	1.653	actinolite	0.77	0.58
n_z	1.651	1.663	tschermakite	0.14	0.30
n_z-n_x	0.018	0.022	glaucophane	0.09	0.12
$2V_x$	71°	85°	$\frac{Mg}{Mg + Fe^{II}}$	0.66	0.52
Z<c	21°	22°			
			$\frac{Fe^{III}}{Fe^{III} + Al^6}$	0.72	0.43

Gabbro in the subgreenschist facies and in the chlorite zone of the greenschist facies has been deformed by fracturing and slips on closely spaced slickensided fault planes; no intragranular or intergranular flow took place and minerals are undeformed and randomly oriented. Effects of intragranular deformation are first visible in gabbro at the biotite isograd: amphiboles and plagioclase are slightly bent and plagioclase simultaneously recrystallized to a granular fabric of albite-epidote. The ophitic texture is still visible at this stage. At the garnet isograd these metagabbros were transformed to amphibolites composed of granoblastic albite and epidote and nematoblastic amphibole oriented with (010) parallel to the schistosity. At this stage generally no trace of the original ophitic texture remains.

Hydration reactions

Three hydration reactions play a major role in low grade metamorphism in the Labrador Trough:

(1) replacement of pyroxene by chlorite and actinolite,

(2) serpentinization of olivine, and

(3) sericitization of feldspar.

All three reactions require water which was apparently supplied by dehydration of pelitic sediments.

Hydration of gabbroic sheets proceeded from their contacts with intercalated shales, and from faults. Hydration is incomplete in the lowermost part of the greenschist facies, as evidenced by relict pyroxene in the central part of the sills. However, intercalated pelite apparently supplied sufficient water for complete replacement of pyroxene by actinolite in the upper part of the chlorite zone. Serpentinization of peridotite consumes much more water than the replacement of pyroxene by amphibole, therefore much of the magmatic olivine persists in the central part of peridotite sills, even in the biotite zone. The intensity of the sericitization of feldspar in the basement gneiss decreases away from the contact with the overlying Proterozoic sequence.

Amphibolite and Granulite Facies

Mineral associations typical of the amphibolite and granulite facies are shown in Table 6. The first appearance of oligoclase (+ epidote) instead of the association albite-epidote defines the lower limit of the amphibolite facies, but the plagioclase isograd has not commonly been mapped. The plagioclase isograd is located between the isograds defined by the first appearances of almandine and of staurolite in pelite, and approximately coincides with the isograd defined by the first appearance of almandine and cummingtonite in iron-rich mafic rocks. Staurolite, kyanite, and almandine are typical minerals, whereas cordierite has been observed at only one locality and andalusite is absent.

The amphibolite facies has been subdivided by several isograds based on the mineral associations in pelitic rocks (Fig. 4). This subdivision is mapped with some difficulty since

Table 5

Overprint of primary textures during metamorphism

Medium to coarse grained mafic rocks (ophitic texture)	Subgreenschist Facies: (1) Replacement of orthopyroxene and olivine by chlorite; (2) Replacement of plagioclase by albite, clouded by very fine grained epidote and/or pumpellyite; (3) Chlorite-filled corrosion canals in plagioclase; (4) Local replacement of plagioclase by prehnite.
	Chlorite Zone: (1) Replacement (partial or complete) of clinopyroxene by actinolitic hornblende I with parallel b and c axes; twins, partings, and lamellae of chloritized orthopyroxene are preserved. (2) Overgrowth of fibrous hornblende III on actinolitic hornblende I; (3) Reaction of actinolitic hornblende with albite results in blue-green amphibole rim. (4) Replacement of clinopyroxene remaining after (1) by small thick-prismatic crystals of actinolitic hornblende II or by chlorite.
	Boundary to Biotite Zone: (1) Recrystallization of albite + epidote to granular crystals ± preserving original plagioclase shapes. (2) Slight deformation and recrystallization of amphibole I; (3) Recrystallization of amphiboles II and III.
	Biotite and Garnet Zone: Intragranular deformation and recrystallization. Granoblastic feldspar. Nematoblastic amphibole oriented with (010) parallel to the schistosity. Complete and rapid destruction of original texture.
Pelite	Subgreenschist Facies and lowermost Chlorite Zone: Discrete cleavage planes, except in strongly deformed zones. Metamorphic differentiation parallel S.
	Upper part of Chlorite Zone: One pervasive cleavage S_1, discrete cleavage planes parallel to axial planes of F_1-folds. Grain size coarse enough to give phyllitic lustre. Locally metamorphic differentiation parallel later S-planes.
	Biotite Zone: Coarsening of fabric.
Orthoquartzite (silica-cemented)	Subgreenschist Facies: Quartz somewhat strained but sedimentary texture preserved except in strongly deformed zones.
	Greenschist Facies: Strained and recrystallized quartz. Clastic grain shapes not preserved.
Arkose	Subgreenschist Facies and Chlorite Zone: Sedimentary fabric well preserved except in strongly deformed zones.
	Biotite Zone: Strong sericitization of feldspar.
Dolomite Dolomitic sandstone	Gradual increase of recrystallization. Most sedimentary textures and structures except bedding generally destroyed in greenschist facies rocks.

Table 6

Typical mineral associations of the amphibolite and granulite facies (after Gélinas, 1965)

	Kyanite-staurolite subfacies	Sillimanite-muscovite subfacies	Sillimanite-potash feldspar subfacies	Granulite facies
Pelite	Quartz - plagioclase - muscovite - biotite ± garnet ± staurolite ± kyanite	Quartz - plagioclase - muscovite - biotite ± garnet ± staurolite ± sillimanite ± kyanite[1]	Quartz - plagioclase - biotite - microcline ± garnet ± sillimanite	Quartz - plagioclase - biotite - orthoclase ± garnet ± sillimanite
Marbles Calc-silicate rocks	Calcite, dolomite, quartz, diopside, tremolite, phlogopite, scapolite			Calcite - diopside - olivine - clinohumite; calcite - scapolite - epidote - plagioclase
Mafic rocks	Quartz - plagioclase ± epidote - hornblende ± biotite ± diopside Quartz - plagioclase ± epidote - hornblende ± biotite ± garnet Quartz - plagioclase - hornblende ± cummingtonite ± garnet ± biotite Plagioclase ± biotite ± cummingtonite ± anthophyllite			Quartz - plagioclase - hornblende - biotite ± diopside ± hypersthene

[1]Kyanite is unstable relict.

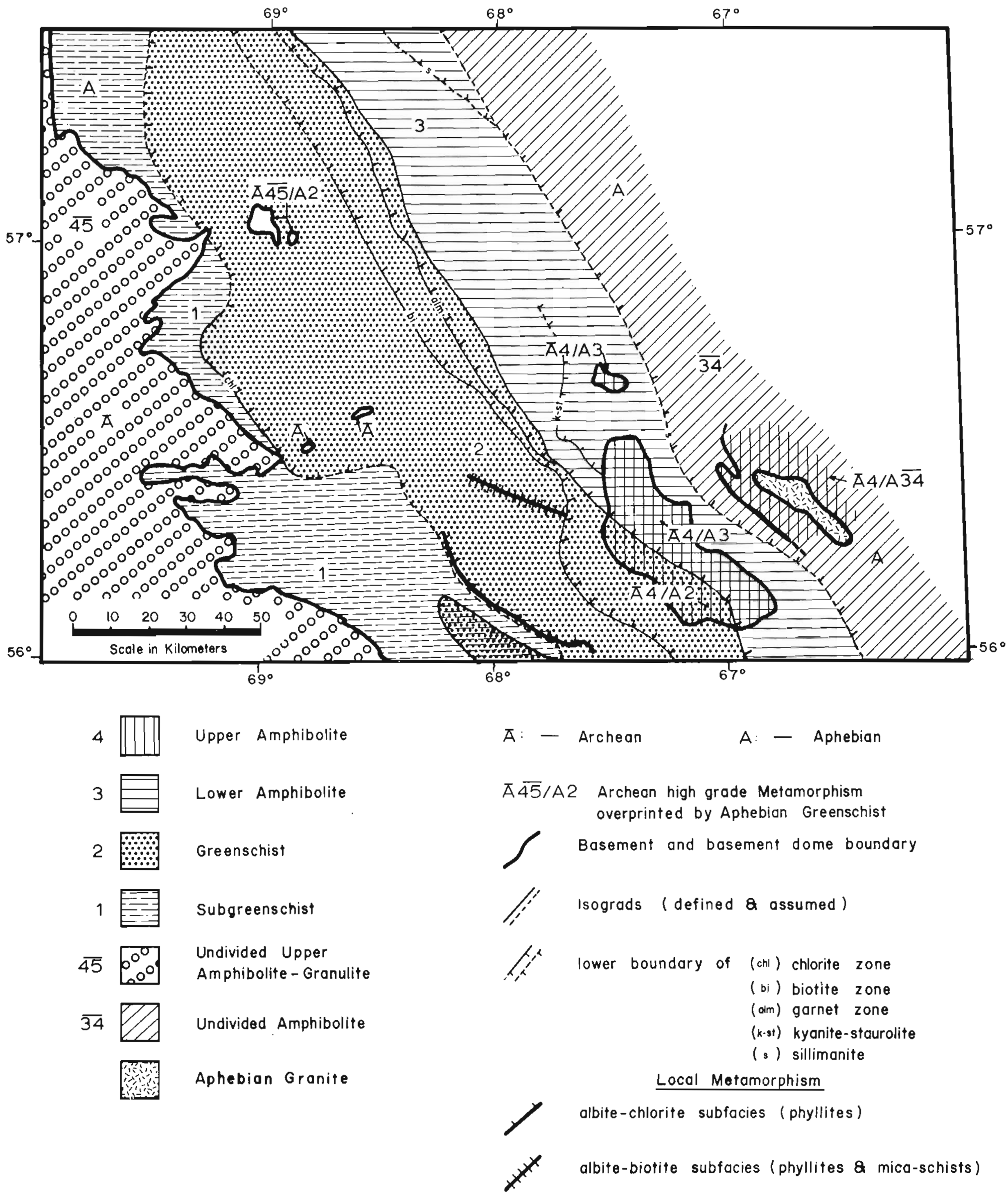


Figure 4. Metamorphic map of the Labrador Trough between 56°N and 57°N.

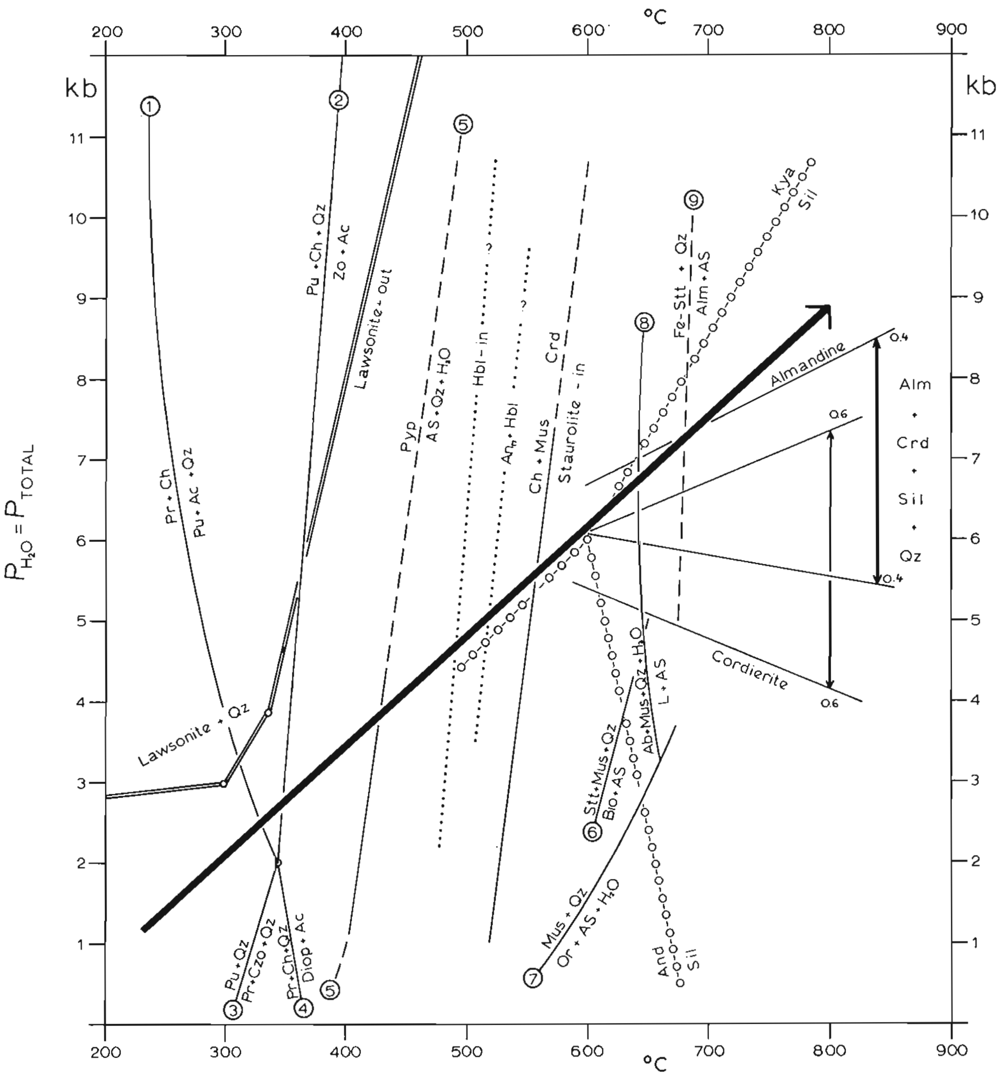


Figure 5

Composite temperature-pressure diagram (after Winkler, 1974) showing petrogenetically important reactions recognized in the Labrador Trough. Path of Labrador Trough metamorphism indicated by bold arrow.

aluminous pelitic sediments are not common. Isograds are defined by the first appearance of kyanite and staurolite, of sillimanite with muscovite, and of sillimanite with potash feldspar. This last isograd coincides with the appearance of widespread anatectic pegmatite veins in the biotite-plagioclase paragneiss, and thus probably corresponds to the reaction

albite + muscovite + quartz = sillimanite + liquid.

Staurolite and kyanite persist well beyond the first appearance of sillimanite. Kyanite is older than the associated sillimanite and is probably a metastable relict (Gélinas, 1965). Staurolite, apparently stable, disappears gradually across the sillimanite-muscovite zone. Armoured relicts of staurolite included in quartz are preserved in the sillimanite-potash feldspar zone and disappear at slightly higher metamorphic grades (Gélinas, 1965), possibly by the reaction

staurolite + quartz = almandine + sillimanite.

Cordierite has been observed at a single locality, co-existing with almandine, sillimanite, and relict kyanite (Gélinas, 1965, p. 31).

Another feature of interest is the existence of reaction zones at the contact between marbles and silicate-rich rocks (notably quartz veins), described in some detail by Gélinas (1965, p. 140-146).

Appearance of hypersthene in addition to augite and hornblende, defines the lower boundary of the granulite facies. Olivine and clinohumite are typical of calc-silicate rocks. Scapolite, also characteristic of metamorphosed calcareous sediments, is present in both amphibolite and granulite facies (Gélinas, 1965).

Zones of High Grade Metamorphism at Fault Zones

Zones of high grade metamorphism accompany some of the larger faults mapped by the senior author. These are generally a few hundred metres wide but may be as much as a kilometre wide. Rocks in these zones are about one subfacies higher than the surrounding rocks. Furthermore, they are strongly tectonized, and are much coarser grained than the rocks of their surroundings. Chown (1976) observed similar high grade metamorphism along fault zones in the Grenville foreland zone.

Table 7

Pressure dependent reactions in Labrador Trough metamorphism

	OBSERVATIONS	INTERPRETATION
1a)	Mafic rocks containing pumpellyite + prehnite + chlorite appear to be followed eastward by rocks containing pumpellyite + actinolite	Reaction prehnite + chlorite = pumpellyite + actinolite + quartz followed by reaction pumpellyite + chlorite + quartz → zoisite + actinolite
1b)	Lawsonite is absent	P = 2 - 3.5 kb at T = 350°C
2a)	Kyanite present, andalusite absent	P = 6 - 7 kb at T = 650°C
2b)	Cordierite absent, almandine present regardless of rock composition	
2c)	Anatexis of biotite-plagioclase paragneiss by reaction albite + muscovite + quartz + H_2O = aluminosilicate + liquid in stability field of sillimanite	

General Conclusions on the Hudsonian Metamorphism

Pressure, Depth and Geothermal Gradients

P-T conditions of the petrogenetically significant reactions in the Labrador Trough are shown in Figure 5 (from Winkler, 1974), and observations bearing on pressure during metamorphism are summarized in Table 7. The observed relations suggest a pressure of 2-3.5 kb[1] at 350°C and of 6-7 kb at 650°C. The first values correspond to a depth of 7.5-13 km at the upper limit of the subgreenschist facies and to a geothermal gradient of 19-43°C/km. The last values correspond to a depth of 22-26 km at the sillimanite-orthoclase isograd and to a geothermal gradient of 25-30°C/km. Both depth values are geologically reasonable only if extensive nappe systems once covered the presently exposed terrane, particularly in the eastern part of the trough. Considering the structure at the exposed level, the former presence of extensive nappes is possible.

Attitudes of Metamorphic Isograds

Folds in the Labrador Trough generally plunge south; successively deeper levels are thus exposed to the north. The western boundary of the greenschist facies zone and the biotite isograd cross the sedimentary and tectonic trend of the Labrador Trough at a small angle (Fig. 1), intersecting the western boundary of the trough at 57°N and 59°N. This is interpreted to indicate shallow westerly dips of both isograds. In contrast, the plagioclase isograd consistently follows the eastern boundary of the volcanic (eugeosynclinal) zone irrespective of the tectonic level; consequently, it is believed that isograds in the amphibolite facies dip steeply (Fig. 9). Since geothermal gradients appear to be comparable throughout the trough, steep dips of amphibolite facies isograds suggest uplift of the amphibolite facies terrane relative to the low grade terrane in the western part of the trough.

Thermal History of the Labrador Trough

Two sets of data permit reconstruction of the thermal history of the Labrador Trough:

(1) Textures, summarized in Table 8 and partly documented in Figures 6-8, permit the determination of time relations between the growth of certain minerals and rock deformation.

(2) K-Ar ages of micas may be interpreted as dating cooling below about 300°C.

Textures suggest that subgreenschist facies conditions were established before the orogeny and were maintained throughout the period of deformation in the western part of the trough. Conditions permitting growth of muscovite and chlorite prevailed through the period of deformation in the centre of the trough. In the east, garnet began to grow at an advanced stage of deformation; kyanite is very late kinematic and sillimanite is postkinematic. In summary, metamorphism in the subgreenschist facies is pre- and synkinematic; lower greenschist facies metamorphism is synkinematic, upper greenschist and lower amphibolite metamorphism is late-kinematic and upper amphibolite facies metamorphism is postkinematic.

On the other hand, K-Ar ages date cooling to <300°C at 1695 Ma in the high grade terrane in the extreme eastern part of the trough, whereas medium grade rocks of the Labrador Trough itself give a cooling age of 1610 Ma (see Dimroth, 1970, p. 2738). Thus, the synkinematic low grade metamorphism in the west apparently was terminated later than the postkinematic high grade metamorphism in the east.

At first sight, these data appear to be in conflict. However, folding generally migrates across a geosyncline from the internal toward the external zones (Aubouin, 1965). Thus, the observed relations may be interpreted as suggesting migration of folding and of metamorphism across the geosyncline, as shown schematically in Figure 10. This interpretation is consistent with other features of Labrador Trough metamorphism and tectonics. The high grade terrane in the extreme eastern part of the trough has been uplifted 10-15 km relative to the low grade terrane to the west. Uplift of such magnitude is inconceivable without considerable deformation. However, the high grade terrane itself must have moved as a block since there is no postcrystalline deformation. It appears logical to assume that part of the tectonic deformation in the western part took place while the highly metamorphosed terrane to the east was uplifted.

Interpretation

Relations outlined above suggest

(1) that metamorphism took place at normal geothermal gradients,

(2) that the easternmost zone of the trough was first covered with a thick pile of nappes, then metamorphosed under near static conditions, and finally uplifted, and

(3) that the deformation and metamorphism in the western part of the trough continued during uplift of the terrane situated to the east.

METAMORPHISM OF THE ARCHEAN BASEMENT

Kenoran Metamorphism

The Archean foreland west of the Labrador Trough is underlain by a monotonous sequence of granitoid gneiss with minor intercalations of amphibolite. Quartz-plagioclase-potash feldspar-hornblende, hornblende-biotite, and biotite gneiss of quartz dioritic and granodioritic composition predominate. The gneiss locally contains garnet or hypersthene. Large granitic masses intrude and/or grade into the gneiss terrane.

[1] 1 kb = 1 x 10^5 kPa.

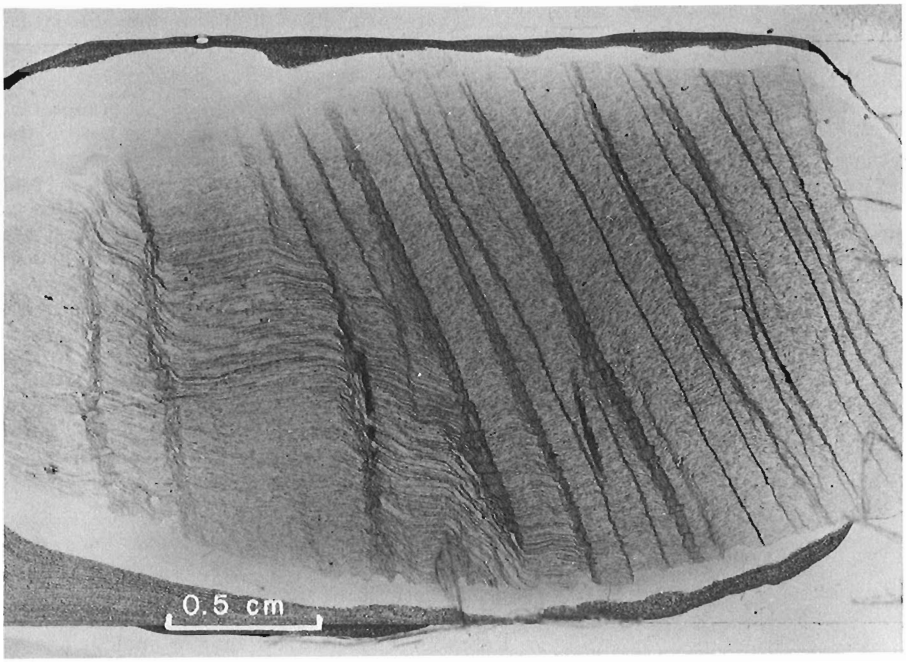


Figure 6A

Quartz-albite-sericite-chlorite schist (upper part of chlorite zone). Note stratification (S_0) and three schistosities (S_1, S_2, S_3). Stratification is poorly visible in thin section, but is underlined by dolomite beds in outcrop. Romanet River. Approximate location 56°20'N, 67°20'W, Section A41-2A. (GSC 203326-D)

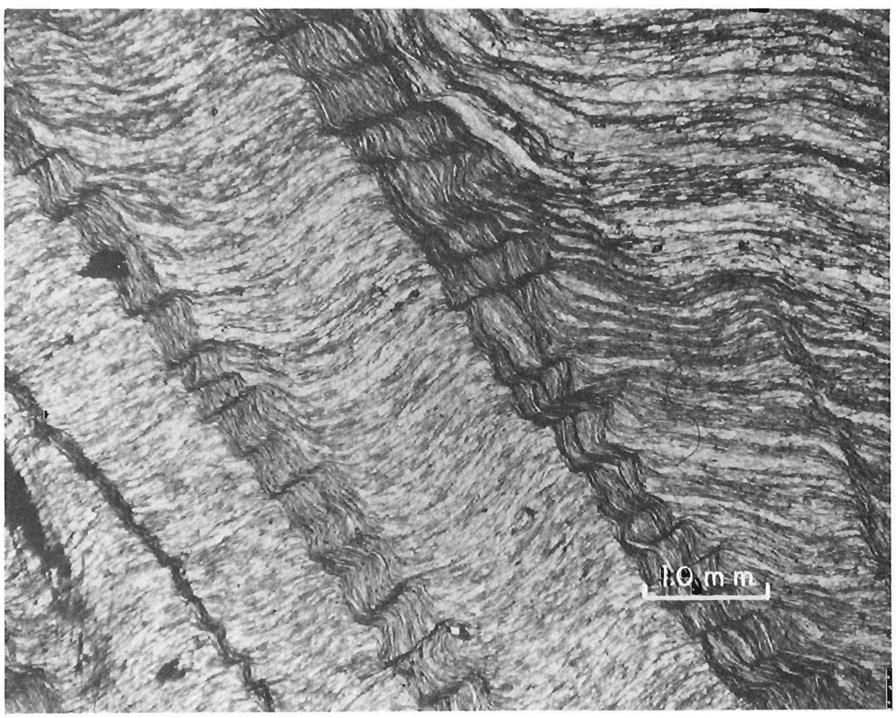


Figure 6B

Detail of Figure 6A. Note the strong metamorphic differentiation, into quartz-albite (sericite-chlorite) microlithons (M) and sericite flaser (S), here parallel to a third schistosity. Microlithons are easily mistaken for bedding; dolomite beds are present in outcrop. (GSC 203326-E)

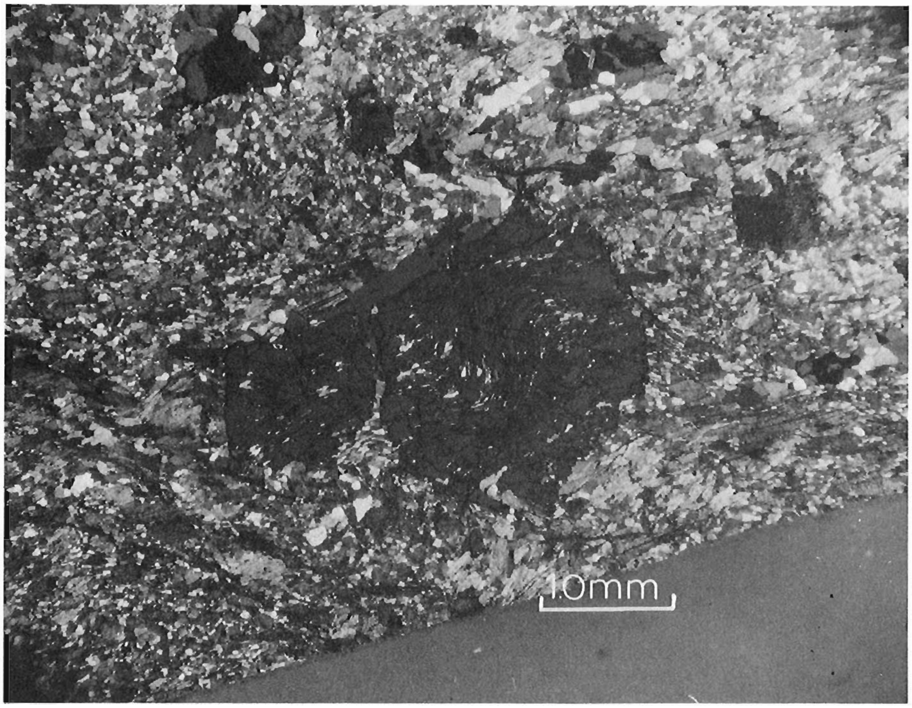


Figure 7

Helicitic texture in garnet. Northwest of Wheeler River. Approximate location 56°25'N, 67°35'W, Section 10-3-8. (GSC 203326-F)

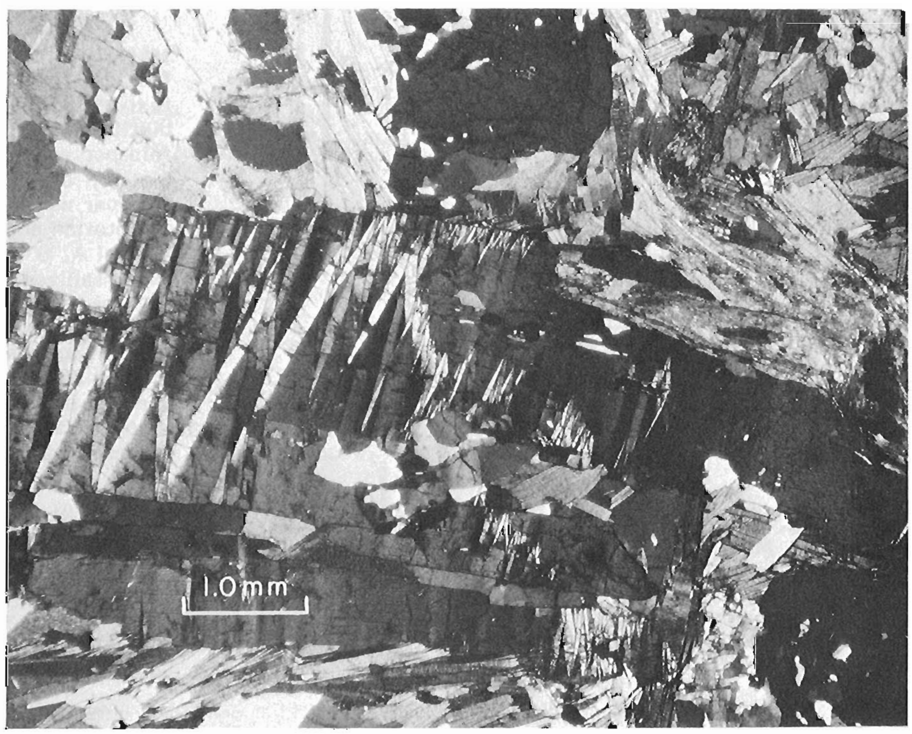


Figure 8

Kyanite-sillimanite gneiss. Note the kink-planes in kyanite (KY). Northeast of Duhamel Lake. Approximate location 56°25'N, 67°05'W, Section 6-4-4. (GSC 203326-G)

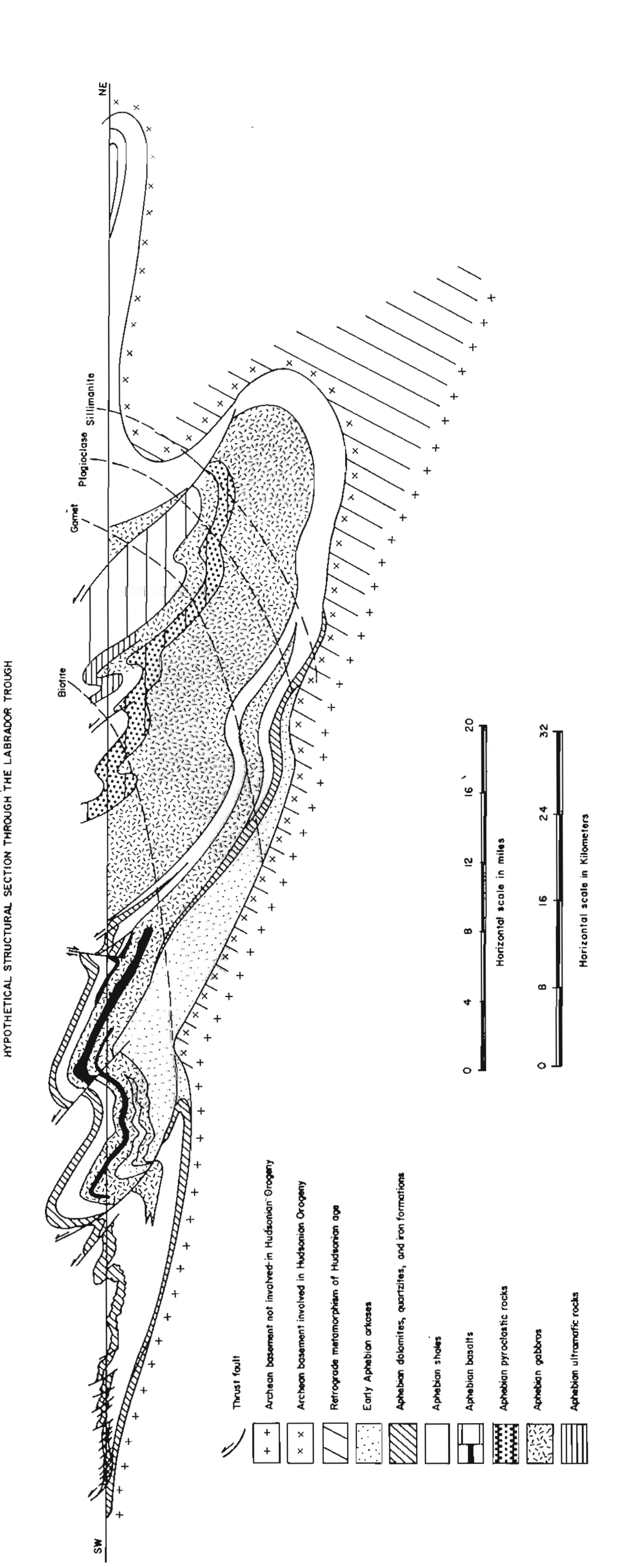


Figure 9. Structural cross-section and attitudes of isograds.

All this gneiss is strongly anatectic. All transitions from metatexites containing well defined granitic, pegmatitic, and aplitic veins, through partly homogenized gneiss with diffuse leucocratic schlieren to essentially homogeneous diatectites are present. Most granitic and granodioritic gneiss probably is the product of anatexis; however there are some cross-cutting granitic dykes and granitic batholiths. The rocks are coarsely granoblastic having been exposed to prolonged static recrystallization at amphibolite or granulite facies conditions.

Hudsonian Retrograde Metamorphism of Archean Gneiss in the Central Labrador Trough

The Archean basement occurs below the Proterozoic sequence in several domes situated between 56°N and 57°N (Fig. 4). During the Hudsonian Orogeny retrograde metamorphism produced similar mineral facies in both the gneiss and the overlying Aphebian rocks. The following descriptions of the remetamorphosed basement gneiss are based largely on work done by the senior author in the Castignon and Romanet Lake areas (see also Table 9).

Subgreenschist and Greenschist Facies

Reactions that affected the Archean gneiss are: transformation of plagioclase to albite + epidote, partial sericitization of the feldspars, and chloritization of biotite and hornblende. In the upper greenschist facies biotite and actinolite formed in place of chlorite.

Subgreenschist and lower greenschist facies metamorphism of Archean gneiss occur only in tight folds and narrow shear zones. The gneiss is generally somewhat cataclastic, with strained and broken feldspars and strained, partly recrystallized quartz. Phyllonitic gneiss similar to gneiss of the upper greenschist facies formed in narrow folds and in shear zones.

Basement gneiss generally had been affected by a severe retrograde metamorphism to the upper greenschist facies. Remnants of the original coarsely granoblastic texture are partly preserved in phyllonitic albite-sericite-biotite gneiss (Fig. 11). They are composed of quartz and feldspar augen in a foliated flaser of fine grained sericite, biotite, albite, epidote, and quartz. Commonly, several feldspar augen have approximately the same crystallographic orientation and may be derived from a single crystal (Fig. 11, crystal A, B and C). Quartz augen, however, are generally recrystallized to a polygonal fabric, so that the derivation of several augen from a single crystal cannot be proven; recrystallization of feldspar augen is less common. Feldspars contain some sericite and epidote. All transitions exist from phyllonitic gneiss to fine grained schistose quartz-albite-sericite-biotite-epidote gneiss phyllonite. Intercalated amphibolite has been transformed to albite-epidote-chlorite-actinolite schist.

Amphibolite Facies

The growth of muscovite at the expense of feldspar and, in the lowermost amphibolite facies, the growth of epidote + sodic plagioclase at the expense of calcic plagioclase, can be recognized. The degree to which original textures have been preserved depends on the intensity of deformation and recrystallization.

Well preserved gneiss retains much of the original fabric, which was defined by stubby lath-shaped plagioclase and interstitial quartz (Fig. 12). Large remnants of the

Table 8

Relations between metamorphism and deformation

Subgreenschist Facies: (1) Minnesotaite and stilpnomelane unoriented in undeformed rocks, are re-oriented into schistosity. (2) Pumpellyite may form lining of pore space and vesicles, but pumpellyite and prehnite also fill tectonic fractures.
Conclusion: Subgreenschist mineralogy stable before and during deformation.
Chlorite Zone: Chlorite and sericite oriented parallel to all S-planes. Development of microlithons associated with early stage of formation of S-planes: Schistosity composed of flasers of sericite, separated by microlithons rich in chlorite and quartz (Fig. 6).
Conclusion: Greenschist facies mineralogy stable during whole deformation.
Garnet Zone: Garnet includes minerals oriented parallel to S_1, but has been rotated (Fig. 7).
Conclusion: Garnet stable at advanced stage of deformation.
Hornblende: Conversion of metagabbro with relict ophitic texture to amphibolite is related to penetrative deformation (Table 5). Hornblende is oriented with (010) parallel to schistosity.
Conclusion: Hornblende formed considerably earlier than termination of deformation.
Kyanite Zone: Kyanite forms porphyroblasts growing across the schistosities but has been kinked (Fig. 8).
Conclusion: Kyanite is very late kinematic.
Sillimanite Zone: Sillimanite forms undeformed porphyroblasts and aggregates.
Conclusion: Sillimanite is post-kinematic.

original plagioclase crystals remain locally (Fig. 12, grain A); in general, most of the feldspar and all of the quartz recrystallized to a fine to medium grained polygonal fabric, and the original texture is defined by aggregates of these polygonal grains. Both feldspars are intergrown with muscovite; plagioclase is associated with epidote in the lowermost amphibolite facies.

Medium grained granoblastic quartz-feldspar-biotite-amphibole-muscovite gneiss are the result of extreme deformation and recrystallization of the Archean basement rocks. They are virtually indistinguishable from biotite-amphibole-muscovite-plagioclase paragneiss derived from the Proterozoic semi-pelitic rocks. However, patches of quartz and feldspar are commonly present and may be interpreted as relicts of the original coarse grained fabric (Fig. 13). Furthermore, the Archean basement gneiss is poorly layered and contains deformed and metamorphosed pegmatitic veins; the Proterozoic paragneiss, on the other hand, generally is well layered and contains intercalations of metaquartzite, calc-silicate quartzite, marble and aluminous gneiss.

Hudsonian Retrograde Metamorphism in other Areas

Little is known about the retrograde metamorphism of Archean rocks outside the area studied by the authors. Wynne-Edwards (1960; 1961) briefly described the retrograde metamorphism of gneiss south of 55°N. Apparently, relations are similar to those observed between 56°N and 57°N. Gneiss in the greenschist facies close to the unconformity appear to have been transformed to phyllonitic gneiss and phyllonite; retrograde effects are less visible farther from the unconformity.

The senior author recognized traces of a retrograde metamorphism in the Archean basement and in diabase dykes having a probable age of 2150 Ma (Fahrig and Wanless, 1963) south of the easternmost part of the Cape Smith-Wakeham Bay Belt. Quartzofeldspathic gneiss has been slightly sericitized, with growth of sericite apparently related to the development of a new schistosity. Furthermore, diabase dykes contain a weak schistosity and have been transformed to actinolite-chlorite schists. Gélinas (1965) and Hardy (1968), on the other hand, did not recognize multiple metamorphism of Archean rocks.

Speculations on the Hudsonian Front North of 59°N

The nature and location of the Hudsonian Front between the Labrador Trough and Cape Smith-Wakeham Bay Belt has always been in doubt. Relevant relations (Fig. 14) are:

(1) the Archean basement gneiss has been involved in Hudsonian folding at the western margin of the Labrador Trough, north of Leaf Bay (59°N), as shown by the configuration of the unconformity.

(2) the Hudsonian biotite isograd intersects the western boundary of the Labrador Trough at 59°N; basement gneiss underlying and to the west of the Aphebian basal units must have experienced temperatures corresponding to the upper greenschist facies north of that latitude during the Hudsonian Orogeny.

(3) Kenoran K-Ar ages are obtained from basement gneiss south of 59°N, whereas the basement gneiss yield Hudsonian ages north of that latitude (Wanless et al., 1968).

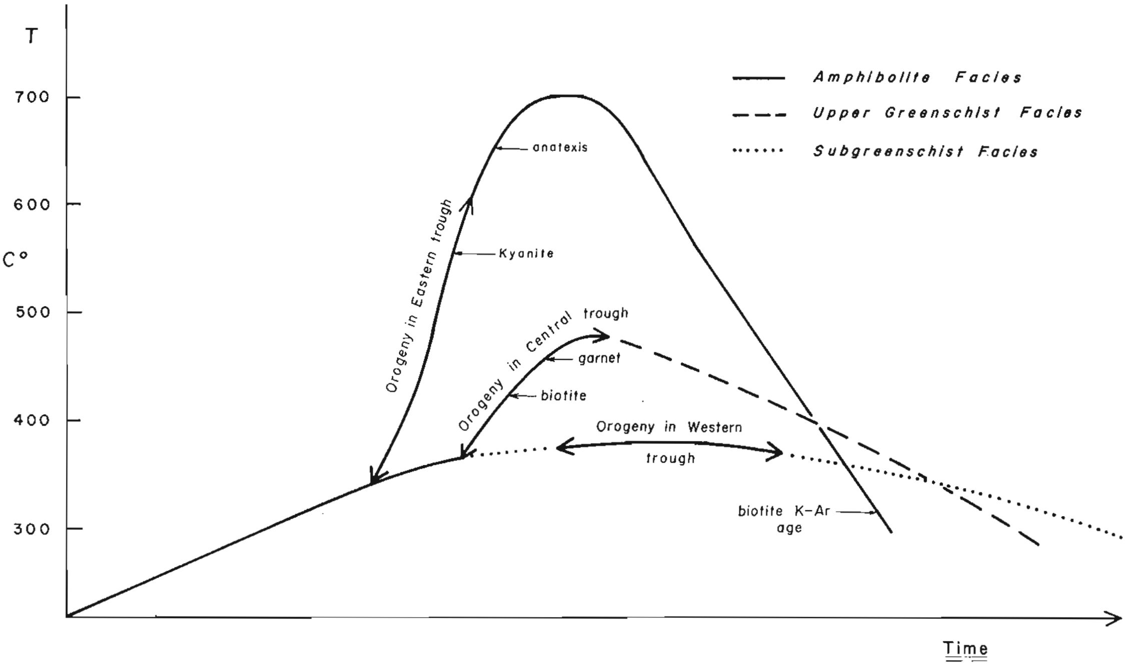


Figure 10. Hypothetical time-temperature curves of subgreenschist facies, upper greenschist facies and upper amphibolite facies terranes.

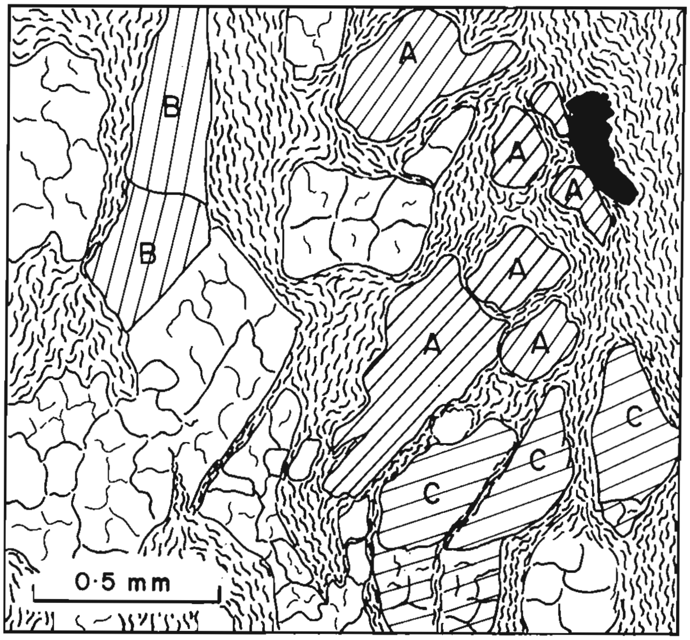


Figure 11. Basement gneiss showing retrograde metamorphism in upper greenschist facies: phyllonitic quartz-albite-epidote-sericite-biotite gneiss. Note remnants of large feldspar crystals (A, B, C) separated by sericite-quartz-albite flaser. The coarse grained texture of original gneiss is well preserved. Wheeler Dome. Approximate location 56°15'N, 67°25'W, Section B45-2.

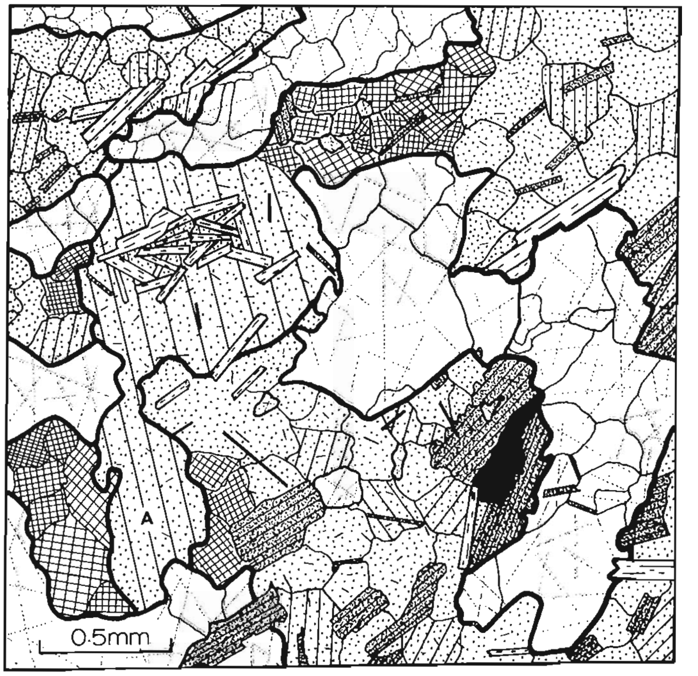


Figure 12. Basement gneiss showing retrograde metamorphism in lower amphibolite facies. Slight deformation with the original coarse grained texture well preserved, but with quartz and feldspar grains recrystallized. Large parts of one original plagioclase grain remain (A). Epidote and muscovite in feldspars. Wheeler Dome. Approximate location 56°25'N, 67°20'W, Section B41-2.

Table 9

Retrograde metamorphism of Archean gneiss

	Subgreenschist Facies	Greenschist Facies	Amphibolite Facies
		Biotite-muscovite subfacies	
Mineral reactions	Feldspar + H_2O = sericite (only in strongly deformed zones).	Plagioclase = albite + epidote. Feldspar + H_2O = sericite Amphibole + H_2O = chlorite related to deformation.	Feldspar + H_2O = muscovite Calcic plagioclase = sodic plagioclase + epidote (lower amphibolite facies).
Typical textures	In strongly deformed zones: Cataclasis of quartz and feldspar, sericite filled feldspar augen, and sericite flaser.	Depending on intensity of deformation: Mortar-, augen or flaser texture. Granulation of quartz and feldspar. Sericite filled relict feldspars and sericite-albite quartz flaser.	Granulation of quartz and feldspar.
Transformation of rocks: Original	Coarsely granoblastic biotite - amphibole gneiss		
Transformation of rocks: Increasing Intensity of Deformation	Cataclastic biotite - amphibole gneiss	Phyllonitic albite - sericite - biotite - (actinolite) gneiss.	Blastomylonitic biotite - amphibole - muscovite gneiss with preserved relict texture.
	Phyllonitic albite - sericite - chlorite - epidote gneiss.	Albite - sericite - biotite phyllonite. Albite - epidote - chlorite- actinolite phyllonite.	Granoblastic biotite - amphibole muscovite gneiss with deformed and metamorphosed pegmatitic veins.

Thus, fronts of Hudsonian deformation, metamorphism, and rejuvenation of K-Ar ages must extend in the basement complex west of the northernmost Labrador Trough, and must join similar fronts south of the Cape Smith-Wakeham Bay Belt. The three fronts do not necessarily coincide and their location is unknown.

POST-APHEBIAN EVENTS

A few stocks of granite and granodiorite intruded the high grade terrane in the extreme eastern part of the Labrador Trough and its hinterland at the close of the Hudsonian Orogeny. Contact aureoles are not present around these stocks. A stock of post-Hudsonian pyroxenite and gabbro is exposed north of Payne Bay and is surrounded by a well-defined hornfels aureole (Hardy, 1968). The Sims quartzite was deposited on the deeply eroded and metamorphosed Hudsonian Orogen during Helikian time in the southern Labrador Trough, and was intruded by the Shabogamo Gabbro. South of the Grenville Front, rocks of the Labrador Trough, their basement, and their Helikian cover, have been involved in renewed deformation and metamorphism during the Grenville Orogeny.

TECTONIC INTERPRETATION OF LABRADOR TROUGH METAMORPHISM

Metamorphism in general is caused either by (1) tectonic subsidence, followed by thermal adjustment and finally, by uplift, or (2) the rise of thermal domes from the lower crust and mantle, generally surrounding intrusive masses. Properties of metamorphic belts accompanying

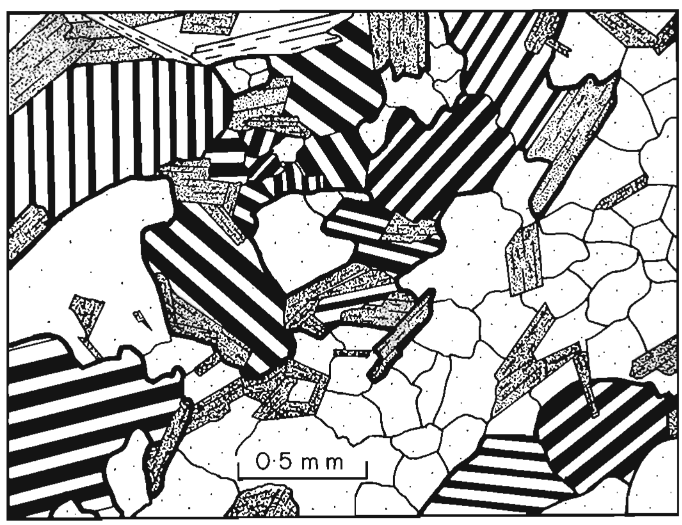


Figure 13. Presumed basement gneiss showing retrograde metamorphism in amphibolite facies. This gneiss is completely recrystallized; however, patches of quartz and feldspar may be remnants of the original coarse-grained fabric. Dome NE of Duhamel Lake. Approximate location 56°29'N, 67°00'W, Section 6-2-4A.

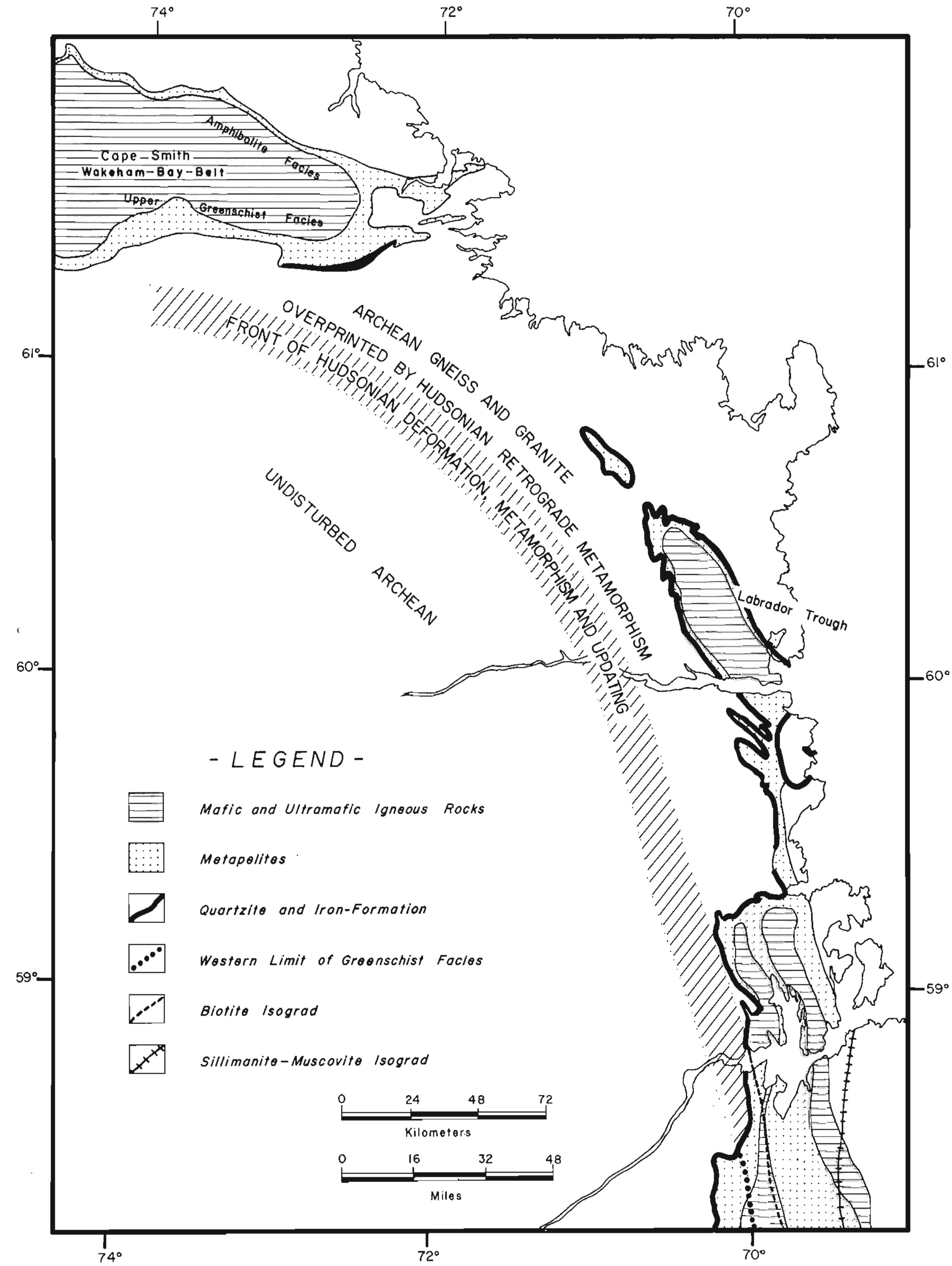
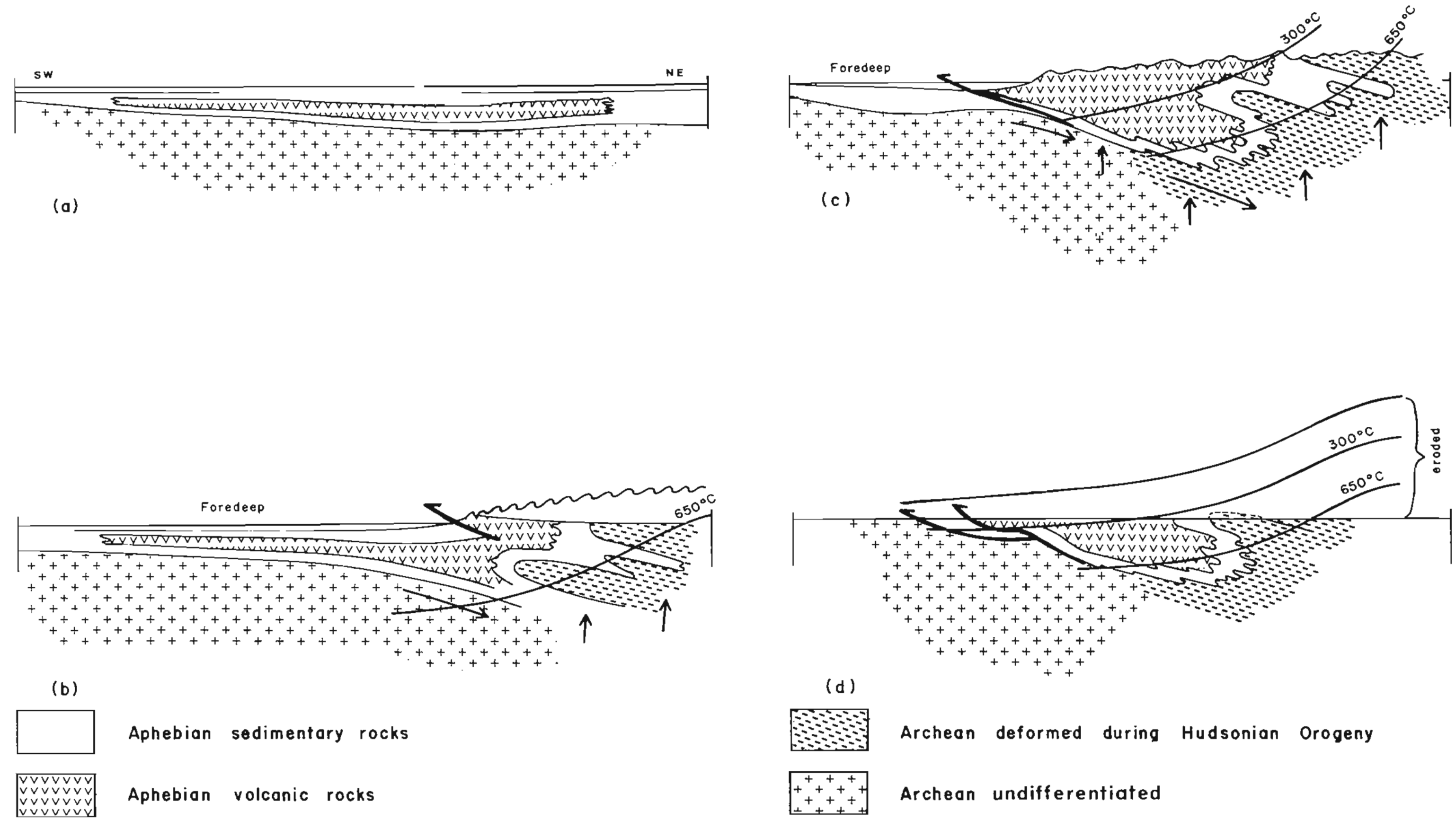


Figure 14. Relations between the northernmost Labrador Trough and the easternmost Cape Smith-Wakeham Bay Belt.



A) Subsidence of the geosyncline

B) Subduction and nappe formation in the extreme eastern part of the geosyncline

C) Beginning of uplift in the extreme eastern part of the geosyncline, continuing subduction farther west

D) Present state

Figure 15. Tectonic and metamorphic evolution of the Labrador Trough.

Table 10

Relations between metamorphic features and tectonics in zones of subduction

A) California, Japan	Paired metamorphic belts: Very high P, low T in the oceanward belt; low P, high T, voluminous granitic masses, extensive calc-alkaline volcanism in landward belt.	Total subduction of thousands of kilometres at rates of 5-10 cm/year. Very incomplete thermal adjustment of subducted plate before uplift. Numerous thermal domes in landward belt.
B) Alpine Belts	High P, low T, but not as extreme as (A), locally overprinted by high P, high T metamorphism (kyanite). Low P belt absent; few granites, little calc-alkaline volcanism.	Total subduction 500-2000 km at rates of 1-3 cm/year. Thermal adjustment of subducted plate before uplift more complete than in (A).
C) Labrador Trough	Metamorphism at normal geothermal gradient. Few granite bodies.	Total subduction 100-200 km (? probably at very low rates?). Complete thermal adjustment of subducted plate.

Cenozoic subduction zones (Miyashiro, 1973; Dewey and Bird, 1970) appear to be related to the rates of subduction (Table 10). Metamorphism also appears to be caused by tectonic subsidence in an outer belt of compressive tectonics and overthrusting (the high pressure belt of Miyashiro, 1973); in this belt, the magnitude of subsidence increases and the rate of thermal adjustment attained before uplift decreases with increasing rate of subduction. An inner belt of metamorphism caused by rising thermal domes (the low pressure belt of Miyashiro, 1973) exists only in areas of very high rates of subduction; it is located in a region where tectonic style is dominated by the rise of batholithic masses and, at high crustal levels, by extensional faulting.

The tectonic style of the Labrador Trough is consistent with a history of underthrusting, although of a somewhat different kind than in the Cenozoic belts referred to above (Dimroth, 1972). In the Cenozoic belts, an oceanic plate has been subducted below a continental area; in the Labrador Trough, a sial-based sedimentary and volcanic basin has been thrust under an equally sialic hinterland. Correspondingly, the Labrador Trough metamorphism stands at the extreme end of the spectrum shown in Table 10: relatively minor underthrusting and subsidence and complete thermal adjustment before uplift lead to the establishment of a normal geothermal gradient after uplift.

Integration of the sedimentary, volcanic, tectonic and metamorphic features of the Labrador Trough suggests the following hypothetical history (Fig. 15):

(1) Subsidence of a wide zone of the Archean crust and deposition of sedimentary and volcanic rocks. Minor extension may have taken place, but an oceanic plate in the presently accepted sense of this term did not form. Load metamorphism occurred in the lower part of the sedimentary and volcanic pile.

(2) Beginning of subduction in the easternmost part of the Labrador Trough and formation of a pile of basement-cored nappes 20-30 km thick in the extreme eastern part of the trough and its immediate hinterland. Sedimentation and volcanism probably continued farther west.

(3) Complete thermal readjustment of the pile of nappes in the extreme eastern part of the trough and metamorphism at normal geothermal gradients, followed by uplift of the highly metamorphosed pile as a block. Continuing subduction, underthrusting and synkinematic low grade metamorphism in the western and central parts of the trough coincided with uplift of the hinterland.

ACKNOWLEDGMENTS

Although this paper is largely based on a compilation of the work of many geologists, it could not have been written without extensive experience in the Labrador Trough. Therefore my first debt of gratitude is to the Quebec Department of Natural Resources which gave both authors the occasion to work in this fascinating region for many years. Bruce L. Penrose kindly supplied an annotated metamorphic map of the Horseshoe Falls area, and L. Gélinas kindly loaned a copy of his unpublished Ph.D. thesis, upon which the account of the high grade metamorphism is based. André Ruest compiled the metamorphic map, drew the figures, and took the photomicrographs. This work has been supported by National Research Council operating grant A9145 and by Geological Survey of Canada EMR research agreement D 13-4-71/75.

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METAMORPHISM IN THE CAPE SMITH-WAKEHAM BAY AREA NORTH OF 61°N, NEW QUEBEC

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Abstract

Metamorphism in the Cape Smith-Wakeham Bay area shows an arcuate pattern that is more or less parallel with the large scale structure of the Circum-Ungava Belt. Aphebian greenschist facies metamorphism affected the main part of the Cape Smith Belt and its Archean foreland. Northerly, towards the hinterland the metamorphic grade increases successively to amphibolite and granulite facies. The granulite facies rocks are in part overprinted by amphibolite facies assemblages. Evidence favours a model involving "early" Aphebian prograde metamorphism and "late" Aphebian retrograde metamorphism.

The successive occurrence of staurolite, kyanite, and sillimanite in the prograde sequence indicates that the Aphebian metamorphism in this area represents an intermediate-pressure facies series. The first occurrence of sapphirine in this area is reported from Cape Hopes Advance.

Résumé

Le métamorphisme de la région du cap Smith et de la baie Wakeham s'est effectué le long d'une trajectoire en arc de cercle, qui est plus ou moins parallèle à la structure de grande dimension de la zone Circum-Ungava. Le métamorphisme aphébien qui dans cette zone a engendré le faciès schistes verts, a affecté la majeure partie de la zone du cap Smith et de son avant-pays archéen. Au nord, en direction de l'arrière-pays, le degré de métamorphisme augmente progressivement, jusqu'à créer les faciès amphibolite et granulite. Les assemblages du faciès amphibolite, ont été repris partiellement dans les roches du faciès granulite. Les renseignements dont on dispose nous amènent à préférer un modèle comportant un métamorphisme prograde aphébien "ancien" et un métamorphisme aphébien rétrograde "tardif".

L'apparition de la staurolite, puis de la cyanite, et enfin de la sillimanite dans la séquence prograde indique que le métamorphisme aphébien représente dans cette région une série de faciès de pression intermédiaire. La saphirine apparaît pour la première fois dans cette région à Cape Hopes Advance.

INTRODUCTION

Reconnaissance mapping by helicopter of the Cape Smith-Wakeham Bay area, north of 61°N was carried out in 1973 by F.C. Taylor, T.M. Gordon, and J.B. Henderson of the Geological Survey of Canada. A short description of the geology and a geological map were published by F.C. Taylor in 1974. Earlier mapping in the area by the Department of Natural Resources of Quebec was concentrated in the eastern half of the Cape Smith Fold Belt. A more detailed geological and petrological study of a part of the Cape Smith Belt (east of 73°00'W) has been done by K. Schimann (in prep.).

The data are based on petrographical analyses of about 550 selected specimens from the Cape Smith-Wakeham Bay collection made available by F.C. Taylor. The specimens were selected from ten of the north-south helicopter traverses, with additional specimens in critical areas.

Acknowledgments

I would like to thank J.E. Reesor, Director of the Regional and Economic Geology Division of the Geological Survey of Canada, for his kind invitation to spend my sabbatical leave, from October 1975 to May 1976, in the Precambrian Subdivision. W.F. Fahrig, Head of the Subdivision, is thanked for his continuous support and encouragement during my visit.

I am very grateful, especially to F.C. Taylor, to be offered the opportunity to study the valuable Cape Smith-Wakeham Bay rock collection. I wish also to thank Dr. Taylor and many other research scientists of the Precambrian Subdivision for the numerous discussions on the complexity and variety of the Precambrian Geology of the Canadian Shield.

GENERAL GEOLOGY

The Cape Smith Belt forms the northernmost part of the Circum-Ungava Geosyncline which surrounds the Archean Ungava Craton of the Superior Province. All three elements of this huge geosynclinal structure, i.e. the Belcher, Cape Smith and Labrador belts, have similar features; the structures are similar and they are composed of the same rock types in a similar succession (Dimroth et al., 1970, p. 114). The Cape Smith Fold Belt consists mainly of Aphebian sedimentary and volcanic rocks. In the south and southeast the basal rocks of the fold belt rest unconformably on an Archean foreland of granitic gneisses. North of the fold belt, the hinterland, of uncertain age, is lithologically similar to the Archean foreland. A complete lithological section through the Cape Smith Belt was studied by Baragar (1974) who described the major stratigraphic divisions as follows:

> "Broadly speaking they are: (1) a lower sedimentary unit of quartzites, dolomites, and shales that rests unconformably on Archean gneisses; (2) a lower volcanic unit comprising massive volcanic flow with interbedded shales and quartzites invaded in part by a profusion of thin (10 to 50 feet) dolerite sills; (3) an upper sedimentary unit composed of quartzites, quartzite breccias and conglomerates and/or shales and minor volcanic breccia and pillow lava; and (4) an upper volcanic unit consisting chiefly of pillowed mafic lavas." (Baragar, 1974, p. 157).

According to Baragar the succeeding units represent facies changes from the basement into the geosynclinal basin.

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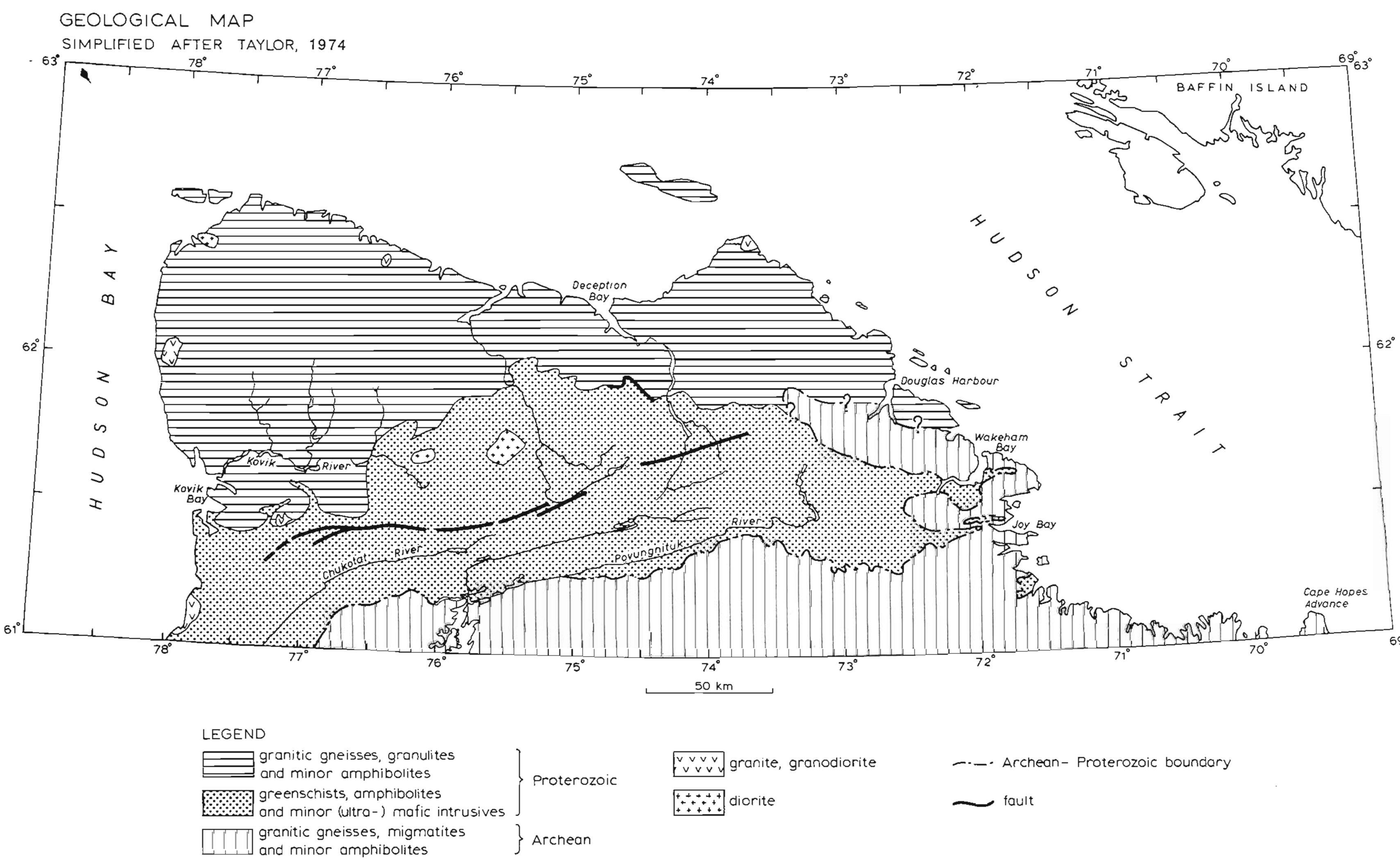

Figure 1. Geological map of the Cape Smith-Wakeham Bay area.

Most of the belt rocks are of a low metamorphic grade. In the central part of the area north of the major fault zone (Fig. 1), however, there are medium grade amphibolites and amphibole gneisses. These rocks are considered to be higher grade metamorphic equivalents of the mafic volcanic rocks.

The Archean basement south and southeast of the Cape Smith Belt is mainly represented by granitic gneisses which appear to have been slightly reworked by Aphebian deformation and metamorphism. The hinterland of the Cape Smith Belt consists of granitic gneisses similar to those in the Archean basement of the foreland. However, amphibolites are more abundant and the occurrence of granulites is a characteristic feature of the hinterland. The present study shows that these granulites form parts of a larger area of granulite facies metamorphism. No reliable age determinations are available for the hinterland rocks. Radiometric data confirm the Kenoran age of the main mineral assemblages of the foreland although overprinting by a younger Aphebian metamorphic event is evident in the gneisses. The ages of the belt rocks fall within the Aphebian Era (Taylor, 1974).

Deformation in the Cape Smith Fold Belt occurred during at least two successive large scale folding phases (Beall, 1960; Taylor, 1974). The older phase is characterized by isoclinal folding with fold axes and axial planes parallel with the overall east-west strike of the belt. A second phase, recognized by Beall (1960), is characterized by more open folding with axes plunging about 45° to the northwest. The surface pattern of these folds displays a sinistral asymmetry. The layering in the Cape Smith Belt dips consistently to the north. Notwithstanding the presence of isoclinal folding, neither Taylor (1974) nor Baragar (1974) favour the theory of tectonic repetition of the strata, mainly because most top determinations of pillow lavas show that the strata are not overturned. The schistosity in the medium to high grade metamorphic hinterland gneisses and amphibolites is much more irregular but retains a predominant northerly dip. The available data on the foreland gneisses suggest at least two sets of foliation, as might be expected in a polyphase deformed terrane (Taylor, 1974).

METAMORPHISM

Polymetamorphism of the Archean basement

The Archean basement rocks comprising chiefly granitic to tonalitic gneisses composed of quartz, plagioclase, biotite and/or hornblende, with or without alkali feldspar, were metamorphosed during the Kenoran Orogeny to amphibolite facies, and later, during the Hudsonian Orogeny, were overprinted by greenschist and amphibolite facies. Structurally below the greenschist-amphibolite boundary, Hudsonian metamorphism has affected less than 10 per cent of the basement rocks. Above this boundary, north of Joy Bay, and in the easternmost part of the area near Cape Hopes Advance, Hudsonian metamorphic recrystallization is probably complete (Fig. 2). Minerals formed near the greenschist-amphibolite facies boundary during the older and the younger metamorphic periods cannot be differentiated. Characteristics of the polymetamorphic mixed gneisses from this zone are variable compositions of plagioclase, and albite rims on oligoclase.

Metamorphic zonation of mainly Hudsonian age

The Cape Smith-Wakeham Bay Area displays a progressive metamorphic zonation characterized by the succession, from south to north, of greenschist facies, amphibolite facies, and granulite facies (Fig. 2). These three major facies zones show a broad arcuate arrangement more or less parallel to the major open arc structure of the Circum-Ungava Belt. Greenschist facies metamorphism has affected the Archean basement gneisses and a major part of the Cape Smith Belt. The boundary between the greenschist and the amphibolite facies is defined by the transition of albite to oligoclase in mafic metavolcanic rock. The oligoclase isograd more or less follows the boundary between the belt rocks and the hinterland gneisses. The greenschist and amphibolite facies assemblages must be of Hudsonian (or younger) age, since they are derived in part from the Aphebian rocks of the Cape Smith Belt. However, no age for the granulite facies metamorphism can be estimated because the age of the rocks that have been affected by this metamorphism is unknown and radiometric age data are not available. The only evidence that might have some bearing on this problem lies in the geological distribution of the granulite facies metamorphism.

Granulite facies rocks occur in irregular masses in the northwestern part of the area. They are defined as stable hypersthene-bearing felsic and mafic assemblages. Vast areas of retrograded granulite facies rocks in the northwestern and the northeastern part of the region and at Cape Hopes Advance suggest that the former extent of the granulite facies zone must have been much greater. Figure 3 shows all known orthopyroxene occurrences. The limit of the orthopyroxene occurrences may represent an original granulite facies boundary which roughly parallels the oligoclase isograd of Hudsonian metamorphism. This suggests that the granulite facies represents the culmination of a "normal", progressive metamorphic zonation of Hudsonian age. By an alternative interpretation the granulite facies metamorphism represents an older (Kenoran) orogenic cycle. This implies that the granulites and probably the main part of the hinterland gneisses are rejuvenated Archean basement rocks. Hence, the orthopyroxene isograd would represent a "retrogradation front" rather than a progressive metamorphic isograd.

The occurrence of two rock bodies, one south of Lac Vanasse (61°50'N, 75°35'W) and one south of Lac Lessard (61°43'N, 76°05'W), within the lower amphibolite facies zone of the Cape Smith Belt possibly favours the second interpretation. These bodies have been mapped as diorites. Specimens of the Lac Vanasse body, however, contain plagioclase-orthopyroxene-clinopyroxene assemblages with a typical metamorphic fabric. Whether these rocks represent a deformed noritic intrusive or true high grade metamorphic rocks in a low amphibolite facies environment should be investigated. In the latter case, they would presumably represent basement fragments, either cupolas or tectonically isolated rock bodies[1]. In conclusion the age of the granulite facies metamorphism is uncertain and the age relations with the Hudsonian amphibolite facies not clearly understood.

Greenschist facies

The two major rock types exhibiting greenschist facies metamorphism are the metabasalts of the Cape Smith Belt and the granitic to tonalitic gneisses of the Archean foreland. The metabasalts are characterized by the assemblage albite-chlorite-actinolite-epidote. Quartz, calcite and muscovite may be present. Sphene is a common accessory mineral. Stilpnomelane is irregularly distributed through the belt rocks; biotite seems to be confined to the southern part of the belt. A tentative biotite isograd is indicated on the metamorphic map (Fig. 2), suggesting that the central part may have been metamorphosed lower in the greenschist facies than the southern part. The amount of metamorphic crystallization is irregular throughout the belt. Relict igneous microstructures and minerals are common, e.g. komatiites with pyroxene showing quenching structures, and chlorite aggregates pseudomorphic after euhedral olivine crystals.

[1] Preliminary results of Kd_{Mg-Fe} determinations for co-existing pyroxenes give similar equilibrium temperatures of 800-850°C for the Lac Vanasse body and the northern granulite applying the Wood and Banno geothermometer (1973).

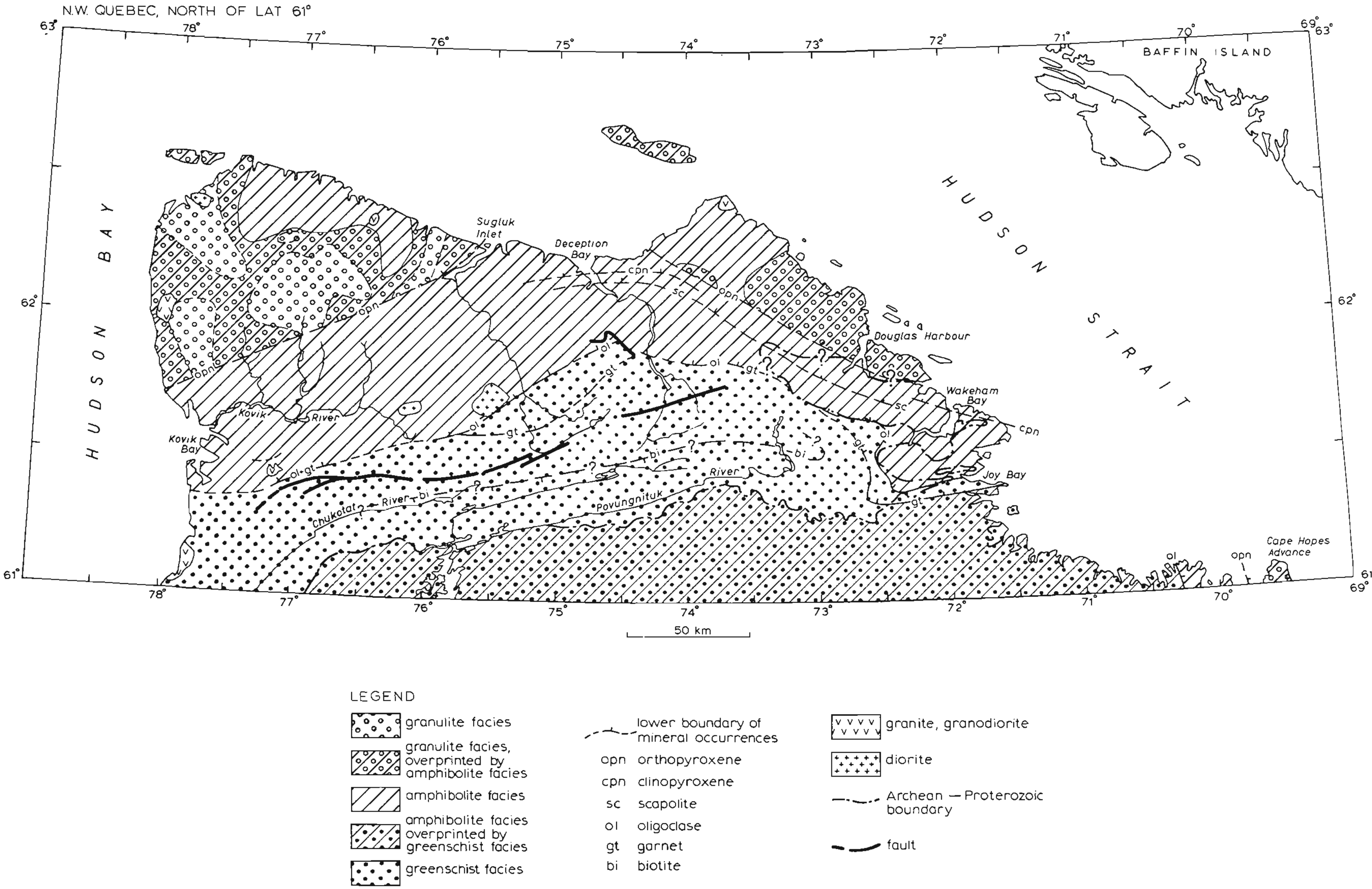

Figure 2. Metamorphic map of the Cape Smith-Wakeham Bay area.

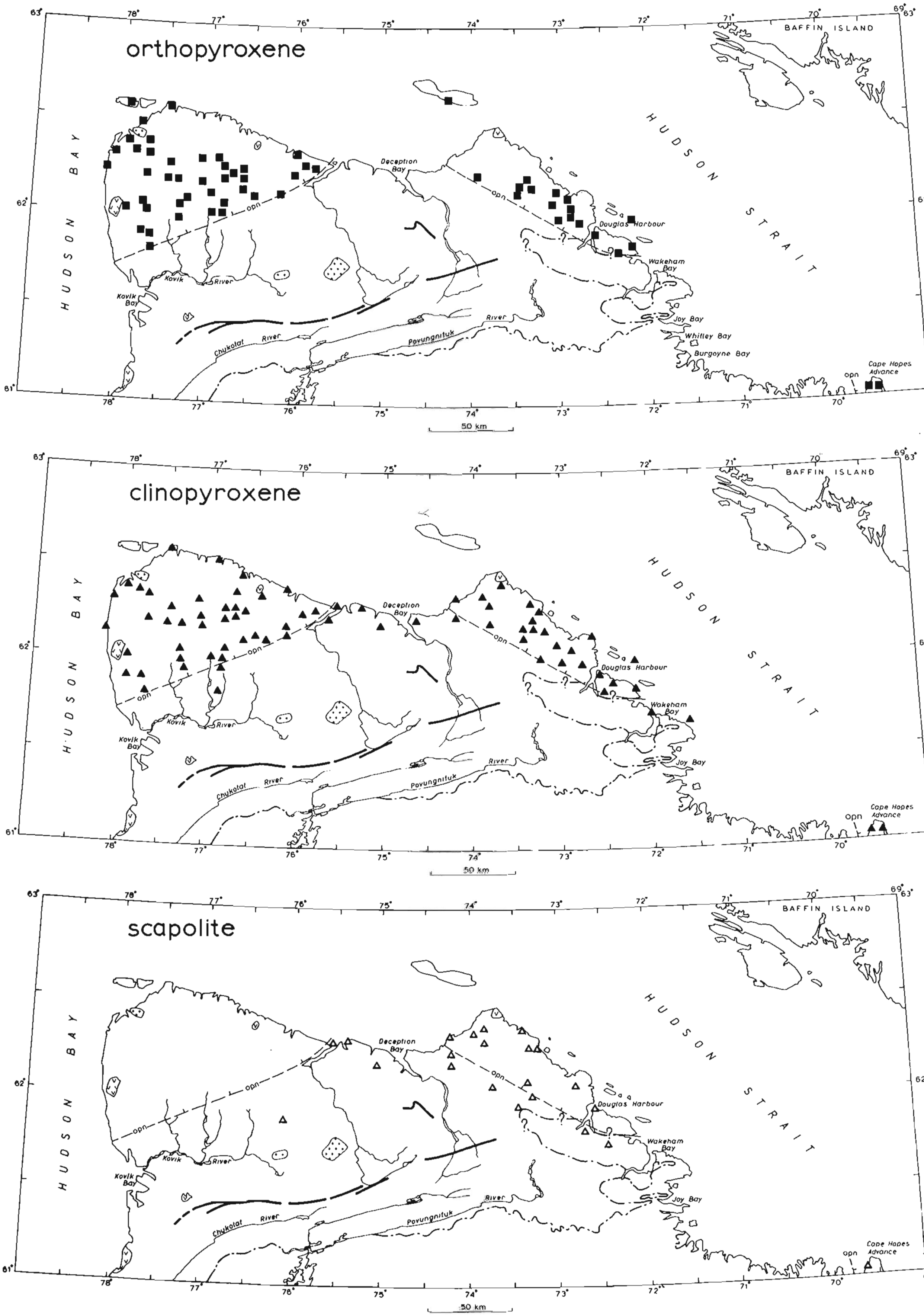

Figure 3. Distribution of high grade metamorphic minerals in the Cape Smith-Wakeham Bay area.

The lower sedimentary unit of the Cape Smith Belt includes iron-formations locally containing grunerite, stilpnomelane, minnesotaite, iron chlorite, and quartz. This assemblage has been studied by Hashimoto and Béland (1968) from an occurrence at Laflamme Lake (61°20'N, 73°45'W). The same unit also contains carbonaceous stilpnomelane-anthophyllite assemblages (Schimann, 1978). The typical pelitic assemblage in this unit is quartz-chlorite-biotite.

The overprinting of the foreland gneisses by greenschist facies is marked by saussuritization and sericitization of plagioclase, recrystallization of quartz and biotite, and possibly by the formation of albite rims on plagioclase. Locally, a newly developed schistosity can be observed. In conclusion, the southern part of the Cape Smith Belt and the foreland gneisses have been metamorphosed during the Hudsonian Orogeny to the biotite zone of the greenschist facies.

A more refined prograde zonation within the greenschist facies of the northern part of the Cape Smith Belt could not be established. Garnet, however, appears locally in the central and easternmost parts of the belt in rocks of lower grade than the albite-oligoclase transition which defines the facies boundary (Schimann, 1978). As the formation of garnet depends on the composition of the metabasalts the limit of garnet occurrences on the metamorphic map (Figs. 2 and 4) cannot be considered as an isograd. The coincidence of the garnet and oligoclase zone boundaries at the northwestern and northeastern parts of the greenschist facies border, and the absence of a biotite zone, may nevertheless indicate a local increase in the metamorphic grade from lower greenschist to lower amphibolite facies within distances of 1 to 2 km.

Another aspect of greenschist facies margin is that it follows precisely the northern and eastern boundary of the Cape Smith Belt (Figs. 2 and 3). Differential heat conductivity and possibly water content between mafic belt rocks and adjacent felsic rocks could explain the narrowing of metamorphic zones at the boundaries of greenstone belts. Hence, in the Cape Smith Belt the greenstone bodies may have caused a steepening of the isograds as well as a shortening of the distance between the isograds during metamorphism, and thus a controlling factor would be the geological time of increased heat flow during the related metamorphic event. In other words, there might be a considerable difference in the time necessary to complete metamorphic reactions in fundamentally different crustal segments.

Amphibolite facies

The amphibolite facies zone is occupied by amphibolites and granitic to tonalitic gneisses. The basic rocks are characterized by the assemblage plagioclase-hornblende. The gneisses contain quartz, plagioclase, alkali feldspar, biotite, and/or hornblende. Both rock types may contain garnet. Pelitic assemblages in the lower sedimentary unit on the coast between Joy Bay and Wakeham Bay give abundant information about the garnet and oligoclase distributions (Fig. 4). Moreover, occurrences of staurolite and Al-silicate minerals indicate a progressive metamorphism to the northeast (Schimann, 1978).

A more refined progressive zonation in the amphibolite facies of the entire area could not be established because of the scarcity of peletic rocks. The few occurrences of staurolite and Al-silicates, however, correspond to a general model in which metamorphic grade increases in a northerly direction. Staurolite occurs closer to the lower boundary of the amphibolite facies, and the Al-silicates farther away from this boundary.

Kyanite and sillimanite occur not only in the amphibolite facies zone sensu stricto, but also in amphibolite facies rocks interpreted as retrograded granulites. Further analyses of rocks from the Geological Survey of Canada collection will probably provide data that will distinguish, on the basis of the occurrence of Al-silicates, between a southern lower amphibolite facies and a northern upper amphibolite facies zone.

The presence of kyanite and sillimanite in the area indicates that progressive metamorphism has taken place along a geothermal gradient above the Al-silicate triple point. Metamorphism in the Cape Smith-Wakeham Bay area is therefore characterized by an intermediate pressure facies series.

Granulite facies

Granulite facies rocks form several isolated irregular shaped bodies in the northwestern part of the area. Typical assemblages contain quartz, plagioclase, and orthopyroxene (probably hypersthene); commonly present are alkali-feldspar, clinopyroxene (probably diopside), garnet, olive green hornblende, and biotite. Felsic granulites in many places contain perthite and antiperthite. The absence of sphene is characteristic. Most of the rocks show a granoblastic microstructure. Platy quartz, and other features of granulite facies metamorphic rocks are common. Hypersthene-bearing assemblages in rocks of varying "charnockitic" composition commonly contain diopside. Diopside, associated with scapolite in many places, also occurs outside the hypersthene areas in rocks characterized by the assemblage diopside-garnet-brown-green hornblende-sphene-scapolite. These assemblages, which represent a Ca-rich rock composition occur in areas where most of the rocks contain assemblages diagnostic of the amphibolite facies. Figure 3 shows the distribution of the high grade minerals orthopyroxene (hypersthene), clinopyroxene (diopside), and scapolite. The distribution of hypersthene has been used to delineate an approximate orthopyroxene isograd, which marks the extent of the granulite facies (Fig. 2). Tentative isograds of scapolite and diopside (Fig. 2) suggest that scapolite and diopside represent successive zones of progressive metamorphism either in the upper amphibolite facies or in a facies transitional from amphibolite to granulite.

At Cape Hopes Advance sapphirine[1] has been identified in one rock specimen. It occurs in a hypersthene-diopside-garnet-kyanite assemblage.

Granulite facies overprinted by amphibolite facies

Most of the hypersthene-bearing assemblages northwest and northeast of the tentative orthopyroxene isograd (Fig. 3) represent granulite facies rocks overprinted by amphibolite facies assemblages. A variety of symplectic intergrowths between pyroxenes, amphiboles, biotite, and quartz give evidence of reactions between the participating phases. Conclusive evidence is present for the following reactions in granulite facies assemblages:

(1) hypersthene → cummingtonite (+quartz ?);

(2) hypersthene → hornblende (+ quartz ?);

(3) hypersthene → biotite (+ quartz + feldspar ?);

(4) diopside → hornblende + quartz

Furthermore, ilmenite often shows a rim of sphene. Cummingtonite is rare, and where present has a hornblende rim. In many places green hornblende replacing hypersthene and/or diopside has a rim of bluish green hornblende. The true granulite bodies in the northwestern hypersthene-bearing area are typically surrounded by a zone of retrograded

[1] Sapphirine occurs with spinel as small inclusions in garnet. A provisional electron-microprobe analysis gives the following composition: 60.6% Al_2O_3, 15.2% MgO, 12.8% SiO_2, 8.7% FeO, 00.45% Cr_2O_3.

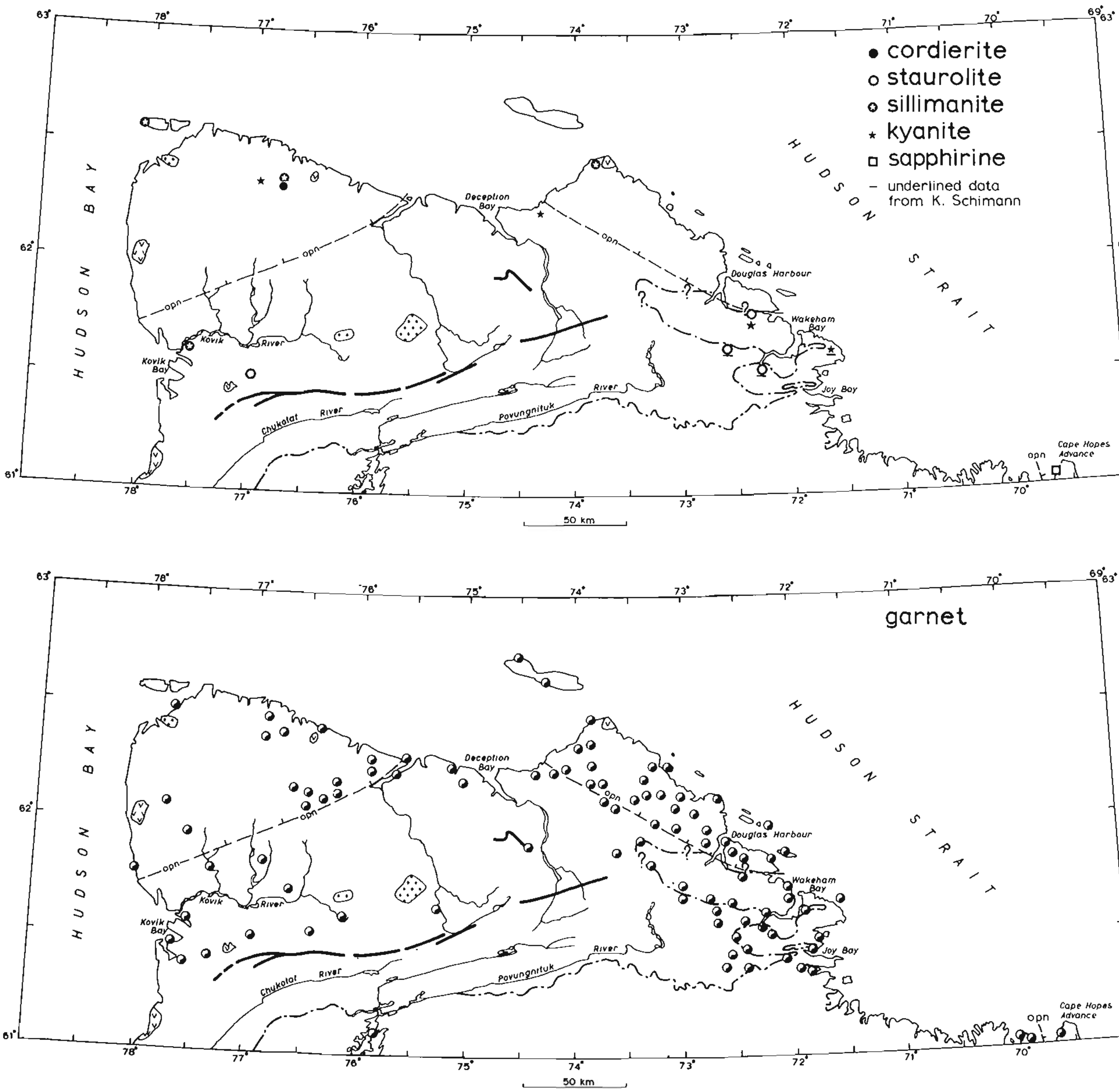

Figure 4. Distribution of medium to high grade metamorphic minerals in the Cape Smith-Wakeham Bay area.

granulites. The reaction products derived from hypersthene and diopside in the retrograded zone increases gradually toward the amphibolite facies terrane that borders the northwest coast where hornblende-quartz symplectites are the only remnants of an original granulite facies metamorphism. Microstructures suggest that a prolonged amphibolite facies metamorphism of older granulites slowly destroyed the originally formed hornblende-quartz symplectites. This overprinting could be described as retrogradation by hydration resulting from a considerable increase in the partial pressure of water but not necessarily accompanied by an appreciable decrease in temperature. There are no true granulite bodies in the northeastern hypersthene-bearing area. The southeast part of this area, around Douglas Harbour, consists of retrograded granulite. In the northwestern part are high grade amphibolite facies assemblages. Only one small body of retrograded granulite, located on the tentative orthopyroxene isograd, has been found in this region.

In conclusion, the features of granulite facies metamorphism and metamorphic zonation in the Cape Smith-Wakeham Bay area can be explained by one of the following models:

(1) The limits of the clinopyroxene and the orthopyroxene distribution define true isograds in a prograded metamorphic terrane that evolved in response to one metamorphic event, probably during the Hudsonian Era.

(2) The granulite facies assemblages are relicts of an older metamorphic period most probably of Kenoran age. This implies an Archean age for these rocks and probably for the main part of the hinterland gneisses.

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ON REGIONAL METAMORPHISM IN THE WAKEHAM BAY AREA, NEW QUEBEC

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Schimann, K., On regional metamorphism in the Wakeham Bay area, New Quebec; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 245-248, 1978.

Abstract

The effects of a high grade Kenoran metamorphism followed by a lower grade Hudsonian metamorphism can be observed in mineral assemblages in the Archean gneisses and in Proterozoic pelitic, basic and ultramafic rocks, and iron formation. The prograde changes in pelitic and basic assemblages, as well as the results of geothermometry (garnet-biotite), and of qualitative (b^0 in potassic white micas) and quantitative (plagioclase-garnet-kyanite-quartz assemblage) geobarometry show the Hudsonian metamorphism to be of intermediate pressure (Barrovian) type, with metamorphic conditions culminating at about 600°C and 6 kb.

Résumé

Les effets d'un métamorphisme kénoréen intense suivi d'un métamorphisme hudsonien plus modéré peuvent s'observer dans les assemblages de minéraux dans les gneiss archéens et dans les roches protérozoique pélitiques, basiques, ultramafiques et ferrifères de la région de la baie Wakeham. Les changements progrades dans les assemblages pélitiques et basiques ainsi que l'application d'un géothermomètre (grenat-biotite), d'un géobaromètre qualitatif (b^0 des micas blancs potassiques) et d'un géobaromètre quantitatif (assemblage plagioclase-grenat-kyanite-quartz) montrent que le métamorphisme hudsonien est d'un type P/T intermédiaire (barrovien) et qu'il a atteint des conditions maximales de l'ordre de 600°C et 6 kb.

INTRODUCTION

The part of the Cape Smith Belt situated east of 73° W (Fig. 1) was mapped for the Department of Natural Resources of Quebec during the summers of 1971 and 1972.

A wide variety of Archean and Proterozoic rocks are present in the area. The Archean basement consists mainly of quartzo-feldspathic gneisses with minor amphibolites and paragneisses. The Proterozoic supracrustal rocks consist of about 80 per cent basic to ultrabasic volcanic and hypabyssal rocks, and about 20 per cent generally immature pelites and psammites, and iron formation.

METAMORPHISM

Two phases of metamorphism can be observed: one, of Kenoran age, affected only the Archean rocks; the other, of Hudsonian age, affected both the Archean and the

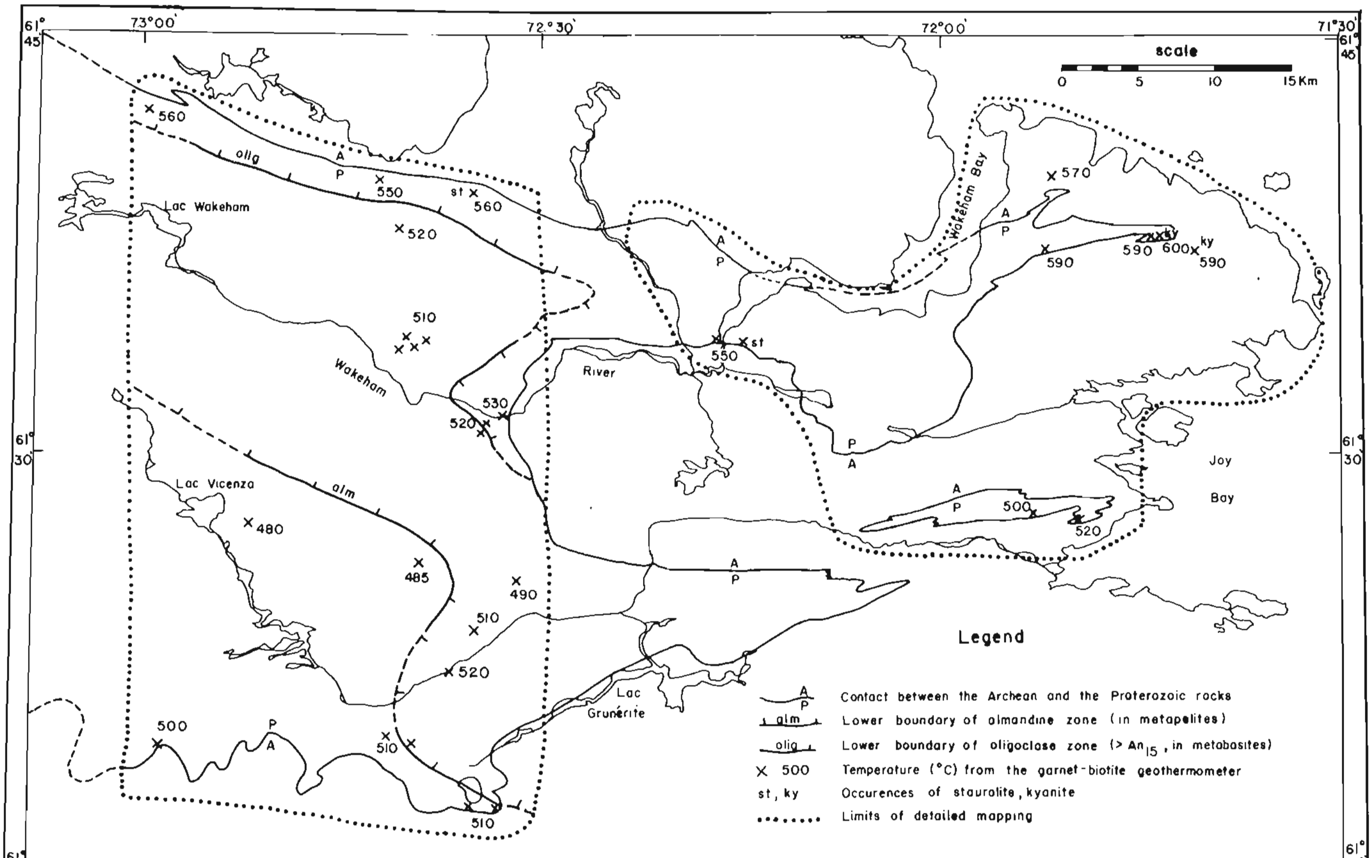

Figure 1. Hudsonian metamorphism in the Wakeham Bay area.

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Proterozoic rocks (Douglas, 1970). The Archean rocks show signs of migmatization (incipient anatexis), indicating that the Kenoran metamorphism probably reached upper amphibolite facies. Only the Hudsonian metamorphism will be considered in this report. This metamorphism increases in grade from south to north across the belt of Proterozoic rocks, and along the belt from west to east. The grade varies from greenschist to mid-amphibolite facies.

Pelitic Rocks

Indicator minerals are rare, owing to lack of rocks of suitable composition; only two occurrences each of staurolite and kyanite have been found. The garnet isograd is based on the presence of almandine with less than 10 per cent MnO. The following assemblages, in order of increasing grade, have been observed:

1) quartz - muscovite - chlorite - biotite - albite
2) quartz - muscovite - chlorite - biotite - garnet - albite
3) quartz - muscovite - magnesian chlorite - biotite - garnet ± staurolite- plagioclase
4) quartz - muscovite - biotite - garnet ± kyanite - plagioclase.

In the lowest grade, corresponding to the biotite zone, K–feldspar is absent, except for a few occurrences as detrital grains in coarse sandstones and in incompletely retrograded Archean gneisses. Stilpnomelane is common in iron formation but rare in metabasites; it is not present in pelites.

Iron Formation

Iron formation is abundant only in the southwestern part of the area (i.e. in the low grade zone)[1], where the following assemblages have been observed:

1) quartz - stilpnomelane - magnetite - carbonate ± grunerite ± biotite
2) quartz - garnet - biotite - chlorite - magnetite ± carbonate ± grunerite
3) quartz - grunerite - magnetite - carbonate
4) quartz - magnetite - hematite.

Magnetite is the common iron oxide; hematite is rare. The chlorite is an iron-rich variety. The carbonate is a manganese- and iron-rich calcite that may coexist with ankerite and possibly siderite. Garnet and stilpnomelane are apparently mutually exclusive. At higher grade, stilpnomelane disappears and the following assemblage is present:

5) quartz - grunerite - magnetite ± garnet ± carbonate ± biotite.

Basic Rocks

Basic rocks are the most abundant as well as the most widely distributed Proterozoic rocks in the area. The low grade assemblage is albite - actinolite - chlorite - epidote - sphene ± quartz ± calcite ± sulphides. With increasing grade, the following changes are observed:

1) actinolite → actinolite + blue-green hornblende → green hornblende
2) albite → oligoclase → andesine
3) sphene is replaced, partly or totally, by ilmenite
4) chlorite becomes more magnesian and finally disappears
5) common epidote, or in places allanite, becomes less abundant, but persists sporadically into the higher grade.

The albite-oligoclase transition delineates the boundary between the greenschist and the amphibolite facies. This boundary coincides also with the following changes: actinolite + hornblende → hornblende; sphene → ilmenite (± sphene). Chlorite is common up to the oligoclase - andesine transition. Garnet, strongly dependent on the Fe/Mg ratio of the host rock, occurs here and there throughout the amphibolite facies; one occurrence was observed in the greenschist facies.

Ultramafic Rocks

Ultramafic rocks in the area are unevenly distributed. The assemblages observed can be classified as low grade (garnet zone of the greenschist facies), medium grade (lower amphibolite facies) and high grade (mid-amphibolite facies). Staurolite is associated with medium grade assemblages and kyanite with high grade assemblages.

1) Low grade assemblages:
   - tremolite
   - tremolite - talc
   - tremolite - talc - carbonate
   - tremolite - carbonate
   - talc - carbonate
   - diopside - serpentine - (olivine, enstatite relicts)
   - diopside - serpentine - tremolite ± carbonate - (olivine, enstatite relicts)
   - tremolite - diopside - (olivine relicts)
   - tremolite - diopside ± carbonate
   - tremolite - serpentine - talc - carbonate
   - tremolite - serpentine ± carbonate
2) Medium grade assemblages:
   - anthophyllite - talc - carbonate
   - tremolite - carbonate
   - tremolite - talc
   - tremolite - biotite
3) High grade assemblages:
   - tremolite - anthophyllite - enstatite - carbonate
   - tremolite - anthophyllite - carbonate
   - tremolite - carbonate
   - anthophyllite - carbonate
   - anthophyllite - carbonate - hercynite - epidote

Opaque minerals (iron oxides and/or sulphides) are additional phases in all assemblages. Chlorite is present in all low grade assemblages and in some of the medium and high grade assemblages. The carbonate is calcite or ankeritic calcite. The presence of relicts of primary minerals (olivine and orthopyroxene) in the low grade assemblages shows that, at least locally, there was a deficiency of water. The common occurrence of carbonates suggests that HCO_3^- was present in varying amounts in the fluid phase during the initial alteration (metamorphism) of the ultramafic rocks. The large number of low grade assemblages observed may thus be explained by the variability of the fluid phase and by the presence of significant amounts of Al and Fe. Serpentine is present only at low grade as is diopside in H_2O- and CO_2-deficient assemblages. Tremolite is present from low to high

[1] The terms low grade, medium grade, and high grade, are used in a relative sense to approximate the greenschist, lower amphibolite, and mid-amphibolite facies respectively.

Table 1
Empirical scale of the b_0 values in low grade metamorphism of pelitic schists (from Sassi and Scolari, 1974)

Facies series	1	2	3	4	5	6
Mean b_0 values	≅8.990	≅8.995	≅9.010	9.020-9.025	≅ 9.035	≅9.055

1. Low pressure metamorphism (andalusite + cordierite); no chlorite (e.g. Bosost).
2. Low pressure metamorphism (andalusite + cordierite); with chlorite zone (e.g. Hercynian metamorphism in Eastern Alps).
3. Low-intermediate pressure metamorphism (andalusite) with the chlorite → biotite → almandine sequence in the greenschist facies (e.g. New Hampshire).
4. Barrovian metamorphism (e.g. Dalradian Series, Scotland).
5. Barrovian-type metamorphism, with simultaneous first appearance of biotite and almandine (e.g. Otago).
6. Glaucophanitic greenschist facies (e.g. Sanbagawa).

grade, anthophyllite at medium and high grade and talc at low and medium grade. The three groups of assemblages (low, medium and high grade) correspond fairly well with the sequence of metamorphic reactions presented by Evans and Trommsdorff (1970). Wakeham Bay assemblages differ in that they contain carbonates but not forsterite. Winkler (1974) showed that in the system $CaO-MgO-SiO_2-H_2O-CO_2$ at 5 kb* forsterite appears at temperatures in excess of 640°C where the fluid phase contained more than 10 per cent CO_2 (i.e. in excess of the highest temperature estimate for the Hudsonian metamorphism in this area).

GEOTHERMOMETRY

An empirical garnet-biotite geothermometer (Schimann, in prep.) was used to determine the temperature of metamorphism at various localities. This temperature varies from about 480 to about 600°C. The almandine isograd corresponds to about 500°C. The four assemblages of pelitic rocks correspond respectively to the following temperature ranges: 1) 480-500°C; 2) 500-600°C; 3) 540-560°C; 4) 580-600°C. The low grade assemblages of the iron formation correspond to 480-500°C (possibly 510°C). In the ultramafic rocks, the low grade assemblage corresponds to 500-520°C, the medium grade to 540-560°C, and the high grade to 580-600°C. The transition from greenschist to amphibolite facies occurs at about 520°C.

GEOBAROMETRY

In order to better define the P/T character of the metamorphism, the b_0 value of the potassic white micas was measured in lower grade pelitic schists, as proposed by Sassi and Scolari (1974). Figure 2 shows the cumulative frequency curve of the b_0 values of the potassic white micas from 65 samples. The mean b_0 is 9.018 (i.e. slightly higher than that in rocks from northern New Hampshire or from Ryoke). On the empirical scale of baric types of metamorphism (Table 1), the Wakeham Bay area corresponds to a slightly lower P/T domain than that of classical Barrovian metamorphism. Indeed the Hudsonian metamorphism in the Wakeham Bay area is characterized by the presence of staurolite and kyanite, the absence of andalusite and cordierite, the prograde appearance of biotite followed by almandine and hornblende and, in basic rocks, the coexistence first of blue-green hornblende (+ actinolite) with albite, followed by green hornblende with epidote.

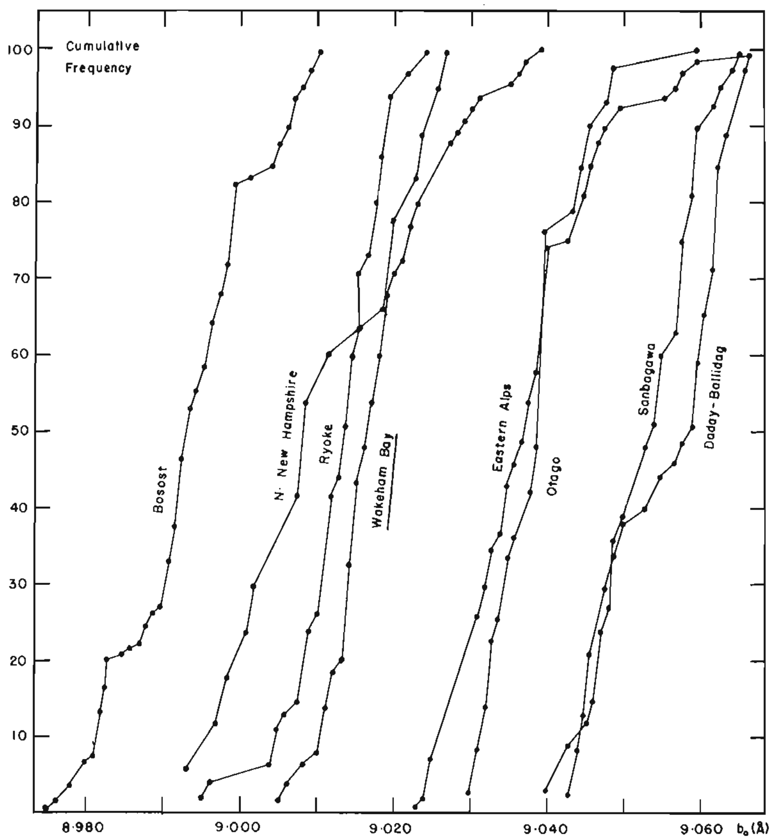

Figure 2. Cumulative frequency curve of the b_0 values of potassic white micas from the Wakeham Bay area compared to those of classical metamorphic areas (after Sassi and Scolari, 1974). Note: The "Eastern Alps" curve corresponds to the Alpine metamorphism in the eastern Alps.

* 1 kb = 10^5 k Pa

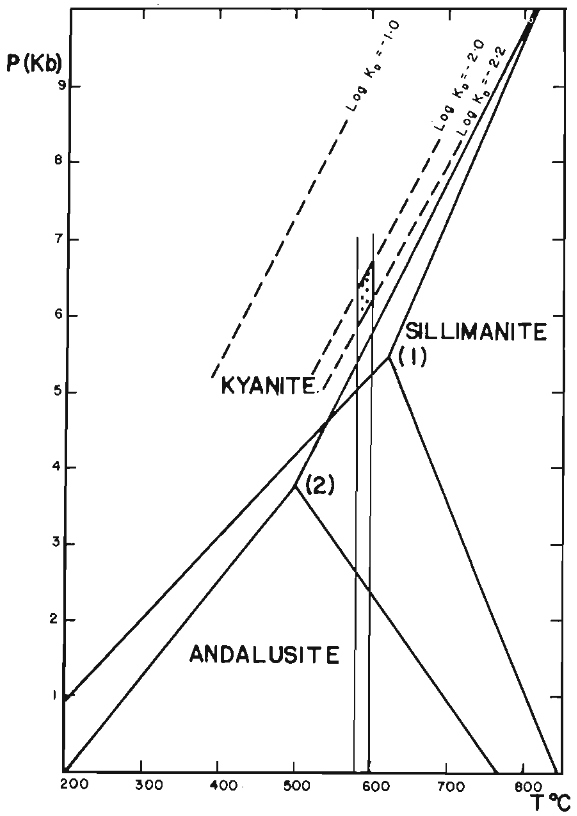

Figure 3. Pressure-temperature diagram showing Al_2SiO_5 phase relationships (1) after Richardson et al. (1969) and (2) after Holdaway (1971). P-T curves for log K_D = -1.00, -2.0, and -2.2 calculated from equation (1b) of Ghent (1976),using an ideal solution model. The stippled area shows the highest grade conditions during the Hudsonian metamorphism in the Wakeham Bay area: 580-600°C, 5.8-6.7 kb.

An estimate of the pressure has also been made for the high grade terrane, using the assemblage plagioclase - garnet - Al_2SiO_5 - quartz. Ghent (1976) showed that the ratio of the mole fractions of grossularite and anorthite in garnet and plagioclase respectively (K_D)* increases with an increase in P/T (i.e. the garnet becomes more calcic). The garnet and plagioclase of the two kyanite-bearing samples were analysed, yielding K_D's of -2.0 and -2.2. The temperature, estimated from the garnet-biotite geothermometer, is 580-600°C. This results in a pressure in the order of 5.8 to 6.7 kb (Fig. 3), closely corresponding to a P/T regime similar to that of Barrovian metamorphism.

CONCLUSION

Two phases of metamorphism can be observed in the Wakeham Bay area: a Kenoran upper amphibolite facies metamorphism and a Hudsonian greenschist to amphibolite facies metamorphism. The effects of Hudsonian metamorphism can be observed both in the Archean basement and in the Proterozoic supracrustal rocks. The boundary between the greenschist and the amphibolite facies (oligoclase isograd in metabasites) parallels the Archean-Proterozoic contact in the northwestern and the central sections, but the eastern half of the area is entirely within the amphibolite facies. Metamorphic temperatures range from 480°C in the southwest to 600°C in the east. Pressure in the high-grade part of the area has been estimated at 5.8 to 6.7 kb. This corresponds to intermediate-pressure type (Barrovian) metamorphism in accordance with a qualitative estimate made on the low grade pelitic schists from the western part of the area, and with the prograde changes in the assemblages of pelitic and of basic rocks.

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* Ghent (1976) defined $K_D = (X_{gr}^{ga})^3 / (X_{an}^{pl})^3$.

PRECAMBRIAN METAMORPHISM ON BAFFIN AND BYLOT ISLANDS

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Jackson, G.D. and Morgan, W.C., Precambrian metamorphism on Baffin and Bylot islands; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 249-267, 1978.

Abstract

Sixty per cent or more of the gneissic terrane on Baffin Island is Archean metamorphites, chiefly granitoid lower to upper amphibolite facies rocks. Some of the Archean rocks were unmetamorphosed prior to deposition of late Archean and Aphebian supracrustal sequences.

Four extensive late Archean to mid-Aphebian supracrustal sequences consisting of shelf, miogeoclinal, and eugeoclinal strata underwent at least one major period of deformation and metamorphism during the Hudsonian Orogeny. Locally, these rocks may have been affected by mid-Aphebian orogenic activity. Metamorphic grade in the late Archean to mid-Aphebian supracrustals is chiefly upper amphibolite and granulite, but ranges to subgreenschist facies. The Hudsonian and possibly mid-Aphebian metamorphism affected mid- and late-Aphebian granite and was also superimposed on Archean basement rocks. Hudsonian metamorphism continued after deformation had largely ceased.

Extensive areas, particularly in central and southeast Baffin Island, are underlain by large intrusive bodies of massive quartz monzonitic and monzocharnockitic rocks. Where both are present, they are separated by a hypersthene isograd. The monzocharnockite is chiefly metamorphosed Aphebian quartz monzonite, but may include unmetamorphosed hypersthene quartz monzonite, recrystallized Archean basement, and recrystallized Aphebian migmatite.

Temperatures as high as 1013°C at an assumed pressure of 10 kb are indicated for co-existing feldspars in granulite facies and monzocharnockite rocks of south-central Baffin Island.

More than 30 km (18 miles) may have been eroded locally from the Baffin Island region since Aphebian time and prior to this the crust may have been as much as 60-70 km (36-42 miles) thick.

Neohelikian strata on northwestern Baffin Island are in the subgreenschist facies. Some Hadrynian diabase dykes may have been metamorphosed to a similar grade, but secondary mineral assemblages are not as diagnostic. Most of the Archean and Aphebian rocks show evidence of diagenetic to possibly greenschist facies retrogression, which may, in part at least, be related to the prograde Neohelikian-Hadrynian metamorphism.

Present metamorphic facies distribution is partly the result of intermittent Neohelikian to Recent faulting and gentle late Proterozoic folding.

Résumé

Au moins soixante pour cent des terrains gneissiques l'île Baffin sont des roches archéennes surtout granitoïdes, dans le métamorphisme s'étend du faciès amphibolite inférieure au faciès amphibolite supérieure. Certaines des ces roches archéennes n'ont été métamorphisées qu'après le dépôt de successions supracrustales de l'Archéen supérieur et de l'Aphébien.

Quatre importantes successions supracrustales, d'âge Archéen supérieur à Aphébien moyen, constituées de strates typiques des plates-formes continentales, miogéosynclinaux et eugéosynclinaux, ont subi au moins un épisode notable de déformation et de métamorphisme pendant l'orogenèse de l'Hudsonien. Localement, ces roches ont sans doute été affectées par les mouvements orogéniques de l'Aphébien moyen. Les roches supracrustales d'âge Archéen supérieur à Aphébien moyen ont été principalement métamorphisées dans le faciès amphibolite supérieure et le faciès granulite, mais aussi dans le faciès schistes verts inférieurs. Le métamorphisme de l'Hudsonien, et peut-être même des épisodes ayant eu lieu à l'Aphébien moyen, ont affecté les granites de l'Aphébien moyen à supérieur, et aussi laissé leur empreinte sur le soubassement. Le métamorphisme de l'Hudsonien s'est poursuivi, même lorsque l'épisode de déformation était pratiquement terminé.

Des zones étendues, surtout au centre et au sud-est de l'île Baffin, contiennent de vastes corps intrusifs composés de monzonite quartzique et de monzocharnockite massives. Là où ces types de roches existent, elles sont séparées par une isograde marquant l'apparition de l'hypersthène. La monzocharnockite est une monzonite quartzique aphébienne en majeure partie métamorphisée, mais peut aussi inclure une monzonite quartzique à hypersthène non métamorphisée, le soubassement archéen recristallisé et une migmatite aphébienne recristallisée.

On a calculé que les feldspaths qui existent à la fois dans le faciès granulite et les monzocharnockites, au sud et au centre de l'île Baffin, ont pu se former à une température de 1 013°C sous une pression supposée de 10 kb.

Depuis l'Aphébien, il est possible que l'érosion ait fait disparaître plus de 30 km (18 milles) de la croûte terrestre dans la région de l'île Baffin, et qu'avant cette époque, la croûte y ait atteint une épaisseur de 60 à 70 km (36 à 44 milles).

Au nord-ouest de l'île Baffin, les strates néohélikiennes ont été métamorphisées dans le faciès sous-schistes verts. Certains dykes de diabase d'âge Hadrynien ont sans doute été métamorphisés au même degré, mais les assemblages minéraux secondaires sont moins typiques. La plupart des roches

[1]Geological Survey of Canada

archéennes et aphébiennes portent les traces d'une rétromorphose, allant de la simple diagénèse au faciès schistes verts; ce métamorphisme est peut-être au moins partiellement apparenté au métamorphisme néohélikien et hadrynien.

La distribution actuelle des faciès métamorphiques résulte partiellement de la fracturation intermittente de la région depuis le Néohélikien et de la phase de plissements légers qui s'est manifestée au Protérozoïque supérieur.

INTRODUCTION

Baffin Island, the largest Canadian island, extends for 1600 km northwest from Hudson Strait to Lancaster Sound. It has an area of 496 400 km^2. Bylot Island (11 180 km^2) is located at the northeast end of Baffin Island (Fig. 1).

Maps at a scale of 1:1 000 000 showing Precambrian metamorphism on Baffin, Bylot, Devon and southeast Ellesmere islands, were prepared for the metamorphic map of the Canadian Shield. The latter two islands are currently being investigated by T. Frisch (Frisch et al., 1978) and are not discussed here. The metamorphic facies were outlined following Winkler (1967, 1974) and the guidelines for the metamorphic map of the Canadian Shield provided by Fraser and Heywood (pers. comm., 1975). The assemblages noted are from thin sections and most are believed to be equilibrium assemblages. No attempt was made to subdivide white mica by X-ray studies and it is reported as muscovite where it seems to form part of the equilibrium assemblage.

This region is chiefly underlain by well-exposed Archean and Aphebian metamorphic and igneous rocks. Slightly metamorphosed Neohelikian strata form much of Borden Peninsula at the northwest end of Baffin Island, and outcrop for about 160 km southeast of the peninsula. Similar strata occur on western and northern Bylot Island, along the south side of Baffin Island north of Melville Peninsula, and in the vicinity of Grant-Suttie Bay. Unmetamorphosed lower Paleozoic rocks outcrop along the western side of Baffin Island from the head of Frobisher Bay northwest to Brodeur Peninsula. Cretaceous-Eocene strata underlie southwestern Bylot Island and outcrop near Cape Dyer, the most easterly part of Baffin Island.

The metamorphic map of Baffin Island is based chiefly on results obtained from four Geological Survey of Canada helicopter reconnaissance bedrock mapping projects led by Blackadar (1967a, 1970b) and by Jackson (1969, 1971; Jackson and Davidson, 1975; Jackson et al., 1975). Results of more detailed investigations of smaller areas have been used to extend facies boundaries between reconnaissance data points and to help interpret the reconnaissance mapping. Reports by Blackadar (1958a, b, 1959, 1962, 1963, 1967b, 1970a), Crawford (1973), Daniels (1956), Davison (1959a, b), Eade (1953), Jackson (1966, 1978a, b), Kranck (1951, 1953, 1955, 1972), Morgan et al. (1975, 1976), Pidgeon and Howie (1975), and Riley (1960) have been especially useful. In preparing the metamorphic map the field notes, field maps, hand specimens and thin sections available from previous Geological Survey studies were re-examined and additional thin sections were made. Over 2500 thin sections and 1500 stained rock slabs were examined.

This paper presents a description of metamorphic features and discusses the nature of metamorphism on Baffin and Bylot islands.

A sequence of events for Baffin Island is summarized in Table 1. The ^{87}Rb decay constant of $1.47 \times 10^{-11} yr^{-1}$ was used to compute Rb-Sr isochron ages.

Acknowledgments

R.G. Blackadar provided field data, thin sections, petrographic descriptions, and compilation maps for northern and southern Baffin Island and for areas bordering northern and eastern Foxe Basin. The geology of Baffin Island and adjacent regions was discussed with R.G. Blackadar, W.L. Davison, T. Frisch, W.W. Heywood, M. Schau, and F.C. Taylor. Advice concerning metamorphism was provided by J.A. Fraser, and by E. Froese. The manuscript was read critically by W. Fahrig and E. Froese. J. Bourne undertook calculations of pressure-temperature fields for co-existing feldspars, and P. Nielsen made similar calculations for biotite-garnet pairs. P. Chernis, C. Kaminini, M.B. Lambert, M. Mazurski, S. Tella and J. van der Leeden assisted in obtaining and compiling data. D. McClintock helped to compile the metamorphic maps of Baffin and Bylot islands.

ARCHEAN METAMORPHISM

A large part of Baffin Island is underlain by rocks displaying some evidence that they were emplaced and first deformed and/or metamorphosed during the Archean. It is considered unlikely, however, that Archean or Aphebian relicts of more than a few square kilometres in area have totally escaped late Aphebian regional metamorphism and deformation.

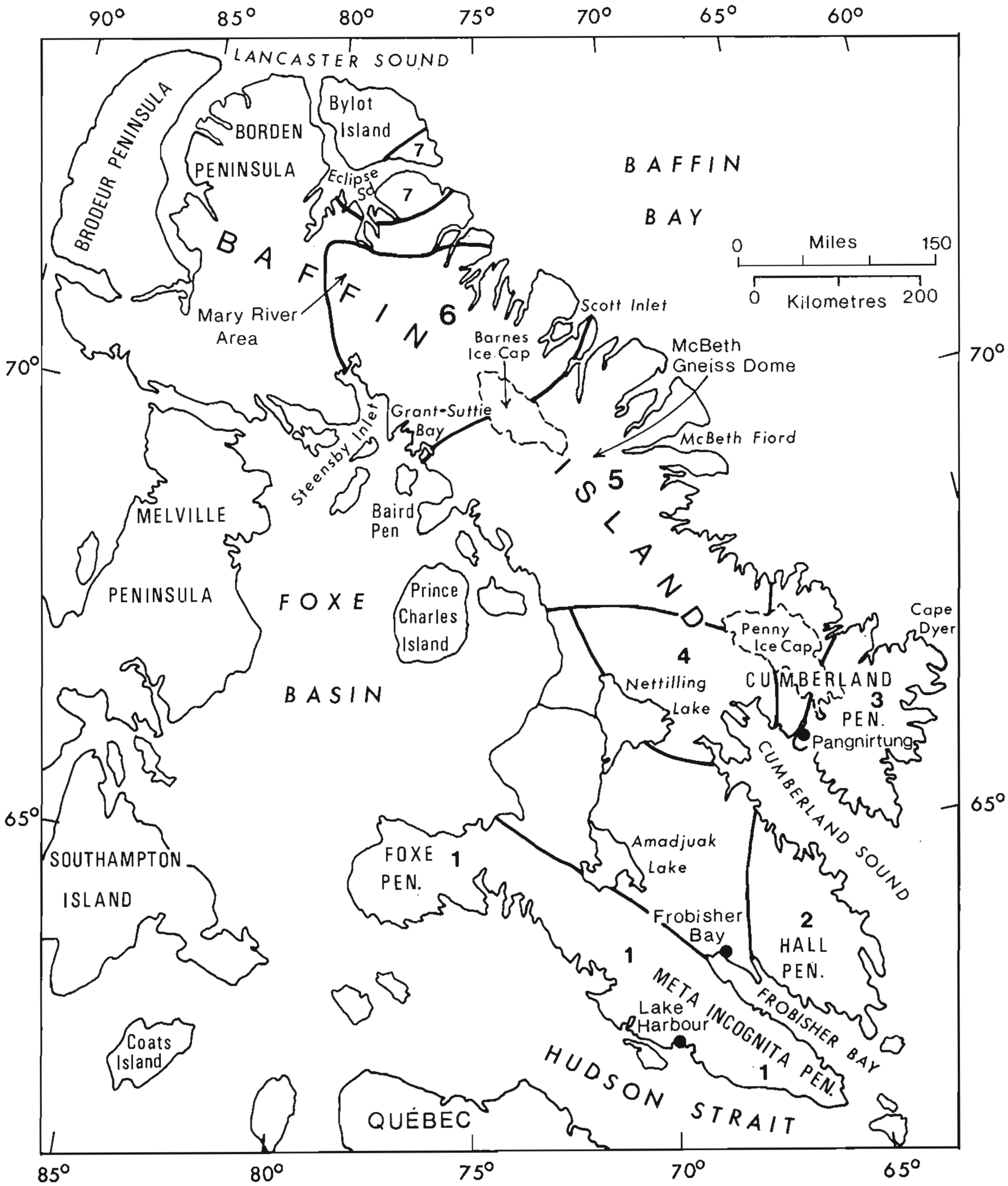
Although a sharp contact between supracrustal rocks and basement gneiss was previously recognized (Eade, 1953; Forrester and Gray, 1967; Gray, 1966; Kranck, 1951, 1953, 1955, 1972), the relationship was interpreted as one between suprastructure and infrastructure; the gneissic basement being regarded as a more highly metamorphosed, deformed equivalent of the supracrustal units. Davison (1959a, pers. comm.), followed by Gross (1966) and Jackson (1966), recognized rocks older than the supracrustal strata. Large areas were interpreted as underlain by Archean rocks as a result of reconnaissance mapping in 1968 and 1970 (Jackson, 1969, 1971; Jackson and Taylor, 1972) and after more detailed field work (Morgan et al., 1975, 1976).

A sharp break with the underlying gneissic granitic rocks is locally well exposed below three supracrustal successions: the Mary River Group, the Piling Group, and the Hoare Bay Group, and may also be present at the base of the Lake Harbour Group (W.L. Davison, pers. comm.). Older granitic rocks that underlie these supracrustal sequences of late Archean and/or Aphebian age, occur in the Mary River area, in the Grant-Suttie Bay area, in central Baffin Island along the northern and eastern margins of the Piling Group, in southeastern Cumberland Peninsula, and in southern Baffin Island. Gneiss domes and basement culminations in these areas are chiefly foliated and lineated granitic rocks and nebulitic granitic gneiss. Structural styles and trends are different and more complicated in the basement gneiss than in adjacent supracrustal strata. The supracrustal rocks near the contact are slightly migmatized only locally and contain preserved sedimentary structures that young away from a basement that commonly shows considerable evidence of anatexis.

Two Rb-Sr isochron age determinations, carried out on nebulitic gneiss and foliated granitoid rocks from the Mary River area and from the McBeth Gneiss Dome (Fig. 1), corroborate the field interpretation that Archean rocks are present in the basement. Samples from the No. 4 Deposit area, near Mary River, have yielded an isochron age of 2552 ± 42 Ma, which is interpreted as a metamorphic age (Jackson, 1978b), and there is an indication that these rocks may be about 2800 Ma old (Wanless and Loveridge, 1978). Samples from the McBeth Gneiss Dome in central Baffin Island gave a Rb-Sr isochron age of 2605 ± 182 Ma. One sample seems to be considerably older than the others (3844 Ma) and a few samples plot along a line that indicates an age of 1964 ± 86 Ma (Wanless and Loveridge, 1978; Jackson, 1978a). The 1964 Ma age and a scattering of points on both isochrons are attributed to an overprinted late Aphebian metamorphism.

Thus in places where there is some recognizable indication of relative age the foliated and nebulitic granitic rocks appear to be relatively old. Using this relationship, we have interpreted these rocks, throughout Baffin Island, as Archean.

At several localities supracrustal strata show no indication of partial melting, in contrast with the adjacent granitic basement rocks which have undergone anatexis, presumably



1 — Lake Harbour Group
2 — Hall Peninsula
3 — Hoare Bay Group
4 — Nettilling Lake — Penny Ice Cap
5 — Piling Group
6 — Mary River Group
7 — Eclipse Sound — Pond Inlet

Figure 1. Regional subdivision of supracrustal strata on Baffin and Bylot islands.

Table 1

Sequence of Precambrian Events, Baffin and Bylot islands

EON	ERA	AGE Ma	EVENT: Intrusive, Extrusive, Sedimentary	EVENT: Tectonic, Metamorphic
PROTEROZOIC	Helikian – Hadrynian	570		Epeirogenic uplift, slight warping, erosion; diagenesis?
		675	SE-trending Franklin tholeiitic diabase dykes	SE tension faults, diagenesis-subgreenschist metamorphism
				SE-tension faults, gentle SE-trending folds
		min 900	Uluksan sediments; Eqalulik sediments, basic volcanics	SE-tension faults, possible aulocogen – NW Baffin Island
				Long period of uplift and erosion. Maximum uplift (18 mi-30 km?) brought granulites to surface. SE-tension faults and cataclased zones.
	Aphebian	1600-1800		HUDSONIAN OROGENY: subgreenschist to granulite metamorphism
		1900-2300	Anorogenic to synorogenic massive and porphyritic granitic intrusions	Deformation and mild metamorphism – possibly local
			Piling, Lake Harbour and Hoare Bay Groups. Shelf, mio- and eugeosynclinal sediments, basic volcanics, basic and ultrabasic intrusions. Acidic volcanics in Hoare Bay Group	
				Uplift and Erosion. Maximum uplift (12 mi-20 km?) brought upper amphibolites to surface
ARCHEAN		2550-2600	Synorogenic? granitic intrusions	KENORAN OROGENY: possibly not developed everywhere, chiefly lower to upper amphibolite metamorphism
			Basic dykes	
		2750	Mary River Group: chiefly eugeosynclinal sediments, acidic and basic volcanics, basic and ultrabasic intrusions, ultrabasic volcanics?	
				Uplift and erosion
			Chiefly NE-trending basic dykes	
		ca 2800?	Synorogenic? granitic intrusions in chiefly granitic terrain. Acidic volcanics	Orogenesis, metamorphism

prior to deposition of the supracrustal rocks. This anatexis reflects late Archean metamorphism which is chiefly of amphibolite grade and areas with these features are shown as undivided amphibolite facies on the Metamorphic Map of the Canadian Shield (Map 1475A, Geol. Surv. Can., 1978).

Locally, conglomerate of the Mary River Group contains acid volcanic clasts of lower metamorphic grade than the containing matrix and associated strata. These clasts are believed to have been derived from nearby Archean volcanic basement rocks (Jackson, 1978b; Jackson et al., in prep.), and their presence is taken to indicate that some Archean rocks may not have been metamorphosed in the Archean or have only attained subgreenschist rank. The widespread granulite facies rocks presently exposed on Baffin and Bylot islands probably were metamorphosed to this grade chiefly in late Aphebian time.

APHEBIAN METAMORPHISM

Introduction

The Aphebian metamorphism affected all Archean and Aphebian rocks on Baffin and Bylot islands and its minimum ages are indicated by the results of K-Ar age determinations. With the exception of two biotite ages, 1865 Ma from the Mary River area and 1975 Ma from Borden Peninsula, the K-Ar ages lie between 1500 Ma and 1800 Ma (Wanless, 1970). Differences in ages in separate regions may indicate regional variations in the timing of post-Hudsonian cooling.

The grade of Aphebian metamorphism ranges from upper subgreenschist to upper granulite facies, with subgreenschist and greenschist facies rocks accounting for less than one per cent of the area. Rocks of amphibolite facies are most abundant in overall extent on Baffin and Bylot Islands, but granulite facies rocks predominate in the southern half of Baffin Island.

Although distinct areas of Archean granitic basement rocks have been recognized adjacent to younger supracrustal sequences (for example Mary River, Piling, Hoare Bay and Lake Harbour Groups), extensive areas of Archean gneissic rocks are intermixed with Aphebian rocks that have undergone high grade Aphebian regional metamorphism. On the metamorphic map these Archean rocks are shown as undivided Archean amphibolite facies overprinted by various grades of Aphebian metamorphism, even though the two metamorphic events have commonly not been differentiated mineralogically or texturally.

Locally at least there is a suggestion from structural, petrological and textural data, as well as from age determinations, that there has been more than one period of Aphebian metamorphism. These different periods of metamorphism have not been discussed separately.

Large areas, particularly in central and southern Baffin Island, are underlain by massive, coarse grained, granitic rocks. This plutonic terrain is considered to be metamorphosed where hypersthene is present and not metamorphosed where that mineral is absent. However, on Map 1475A these massive hypersthene-bearing rocks are shown as areas of granulite facies metamorphism and not as distinctive charnockitic plutons.

Depositional and metamorphic ages of supracrustal sequences

Mary River Group

This group outcrops discontinuously from the Grant-Suttie Bay-Scott Inlet region northwest to Eclipse Sound and Bylot Island (Fig. 1; Jackson, 1978b; Jackson et al., in prep.) and is correlated with the Prince Albert Group on Melville Peninsula (Jackson, 1966) which was considered Archean by Heywood (1967). Zircons from slightly altered acid volcanics from the Mary River Group in the Grant-Suttie Bay area, and from the Prince Albert Group have given U-Pb ages of 2747 ± 15 Ma and 2730 Ma respectively (R.K. Wanless, pers. comm.). These ages are considered to indicate the time of volcanism, and further corroborate the presence of Archean rocks on Baffin Island.

Gneiss basement to the Mary River Group in the Mary River area has given a Rb-Sr isochron of 2552 ± 42 Ma, but these rocks were probably in existence by about 2800 Ma (Wanless and Loveridge, 1978; Jackson, 1978b).

A Rb-Sr isochron, calculated from analyses undertaken by Fryer (1971a, b), indicated a metamorphic age of 1904 ± 94 Ma for the group in the Mary River area (Table 2; Fig. 2; Jackson, 1978b). Points for this isochron were chosen selectively (Jackson, 1978b) and analyses by both Fryer and the Geochronological Laboratories of the Geological Survey of Canada indicate possible ages ranging from 2100 Ma to 2600 Ma, with several in the range 2100-2200 Ma. The 1904 Ma Rb-Sr isochron has an initial ratio of 0.7102 ± 0.004, which is considerably higher than usual for rocks of this age. If the system was closed prior to metamorphism about 2000 Ma ago it would take about 350 Ma to increase the initial ratio from 0.703 to 0.710, and about 450 Ma to increase it from 0.701. This suggests that the rocks in the Mary River area may be younger than those at Grant-Suttie Bay or may have had an abnormally low $^{87}Sr/^{86}Sr$ ratio, or that the system was not closed.

Conglomerate in the Mary River area contains acid volcanic clasts that resemble rocks in the basement to the group. Locally these slightly recrystallized clasts occur in more highly metamorphosed matrices that range from upper subgreenschist to upper amphibolite facies.

The Mary River Group was deposited approximately 2750 Ma ago on a basement that was in part gneissic, but included unmetamorphosed rocks. Deformation and metamorphism about 2550 Ma ago profoundly affected the basement and most of the Mary River Group. Parts of the group probably escaped this late Archean event and were not metamorphosed until late Aphebian time, approximately 1900 Ma ago. The mid-Aphebian dates, about 2100 Ma to 2200 Ma, may represent an event hitherto undocumented on Baffin Island. Highly migmatized Mary River strata are intruded at several localities by white to grey granodiorite which is probably related to the migmatization, but the granodiorite and some of the migmatite may represent a mid-Aphebian event. Late-Aphebian pink quartz monzonite (Jackson, 1966) intrudes both the migmatite and the granodiorite. A Rb-Sr isochron of 2134 Ma has been obtained for gneiss associated with the Penrhyn Group at one locality on Melville Peninsula (R.K. Wanless, pers. comm.). On the basis of current information, which does not permit subdivision of the different periods of metamorphism, mineral assemblages in the Mary River Group are considered to have resulted from late Aphebian metamorphism.

Piling Group

Piling Group strata outcrop northeast from Nettilling Lake and the head of Cumberland Sound to east of Steensby

Table 2

Rb and Sr analyses of Mary River Group samples by B.J. Fryer (1971)

Sample	Rb (ppm)	Sr (ppm)	Rb^{87}/Sr^{86}	Sr^{87}/Sr^{86}
R7804[1,2]	96.8	401	0.700	0.7327
R7805[2]	19.3	149	0.375	0.7196
R7806[2]	5.36	100	0.155	0.7144
R7807[1]	87.9	147	1.74	0.7626
R7808[1]	150	9.62	51.7	2.1700
R7809[1]	245	8.49	108	3.671
R7812[2]	1871	58.3	125	4.1902
R7817[2]	33.4	388	0.249	0.7156

[1] – Metasediment

[2] – Basic Metavolcanic

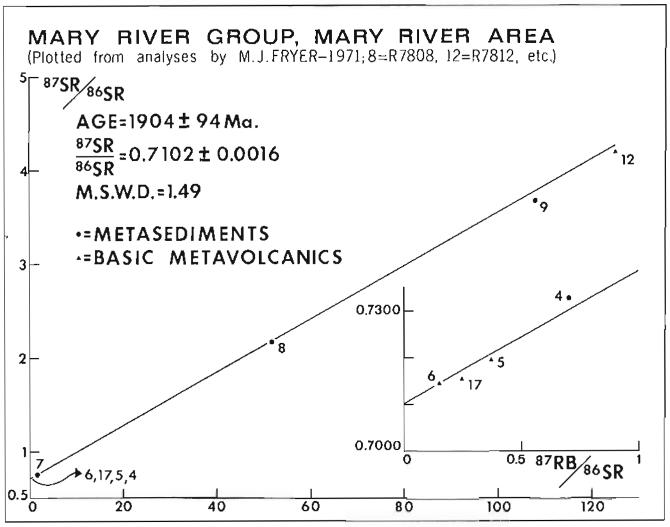


Figure 2. Rb-Sr whole-rock isochron for Mary River Group from data by B.J. Fryer (1971; Table 2).

Inlet and Barnes Ice Cap (Fig. 1; Jackson and Taylor, 1972). The group is correlated with the Penrhyn Group on Melville Peninsula (Jackson, 1969) which was considered Aphebian by Heywood (1967).

Regionally extensive marble, dolomite, quartzite and meta-arkose horizons, such as those contained in the Piling and Penrhyn groups and in other Aphebian sequences, have not been unequivocally documented in Archean sequences. The basal Piling Group arenite unit rests unconformably on basement gneiss and extends east across Baffin Island and south from the Barnes Ice Cap to Penny Ice Cap. The Piling Group carbonate unit occurs chiefly in two lobes: one occupying a triangular area between Steensby Inlet, McBeth Fiord, and Baird Peninsula; the other an area bounded by South Penny Ice Cap, Nettilling Lake, and upper Cumberland Sound.

Only K-Ar age determinations (between 1665 Ma and 1805 Ma) are currently available for the Piling Group and, together with the 1964 Ma overprint indicated by Rb-Sr work on Archean gneiss from McBeth Gneiss dome (Jackson, 1978a), provide a minimum age for the group. The significance of Rb-Sr errorchron ages of about 2300 Ma from selected samples in the McBeth Gneiss Dome is not known; they could conceivably represent the time of inception of the Piling Basin. A Rb-Sr isochron age of 1670 ± 30 Ma (unpublished) has been obtained by the Geological Survey of Canada Geochronological Laboratories for charnockitic granite that intrudes high grade Piling Group strata.

The Piling Group is therefore thought to have been deposited in early Aphebian time and has undergone orogenic deformation and metamorphism in the Aphebian.

Lake Harbour Group

The Lake Harbour Group outcrops throughout southern Baffin Island south and west of Frobisher Bay and Amadjuak Lake (Fig. 1). It is correlated with the early Aphebian Piling and Penrhyn groups and with similar strata in northern Labrador-Ungava and on Southampton Island (Jackson and Taylor, 1972).

K-Ar age determinations for this group and for associated rocks range between 1665 and 1800 Ma. Aphebian deformation and metamorphism of the Lake Harbour Group is indicated by a Rb-Sr isochron of 1812 ± 30 Ma (unpublished). This age has been obtained by the Geological Survey of Canada geochronological unit for banded migmatitic granulites at Frobisher Bay that occur in close proximity to marble and paragneiss lenses.

Hoare Bay Group and Hall Peninsula Strata

Hoare Bay Group rocks outcrop throughout the southern two thirds of Cumberland Peninsula and lithologically resemble the Mary River Group more closely than the Piling and Lake Harbour groups (Jackson and Taylor, 1972). However, quartzite, carbonate rocks, and rusty schist occur in the northern part of the Hoare Bay Group on strike with Piling Group rocks to the northwest and west.

Supracrustal strata on Hall Peninsula are varied but resemble lithologies in the Piling and Lake Harbour groups more closely than in the Hoare Bay Group.

The Hoare Bay Group and Hall Peninsula strata are tentatively considered to be of early Aphebian age, and are regarded as predominantly eugeoclinal facies equivalents of the Lake Harbour and Piling groups. An extension of these eugeoclinal rocks can be traced northwest through the Penny Ice Cap and then west to Foxe Basin south of Baird Peninsula.

Hornblende from massive charnockite north of Pangnirtung and from Hoare Bay amphibolite southeast of Pangnirtung has yielded K-Ar ages of 1721 Ma and 1740 Ma respectively. A U-Pb zircon age of 1900 ± 20 Ma has been reported by Pidgeon and Howie (1975) for charnockitic granulite near Pangnirtung. Therefore, the Hoare Bay Group is also thought to have undergone late Aphebian orogenesis and metamorphism.

Subgreenschist Facies

Rocks containing late Aphebian upper subgreenschist facies mineral assemblages occur discontinuously within the greenschist facies zone in the Mary River area. Typical mineral assemblages include:

(1) Quartz-oligoclase-biotite-muscovite-chlorite-pumpellyite-prehnite-sphene

(2) Quartz-biotite-muscovite-chlorite-epidote-pumpellyite-prehnite

Assemblage (1) is from a metasediment and (2) from a metamorphosed intermediate flow or sill.

Mafic minerals and plagioclase in associated higher grade rocks are downgraded and partially recrystallized to upper subgreenschist facies minerals. Because the Mary River area represents a domain of relatively low grade metamorphism enclosed within a higher grade terrane, the subgreenschist facies metamorphism is interpreted as a closing phase of the late Aphebian metamorphism. A late Archean, or a late Proterozoic age for this event, however, cannot be ruled out (see section on Helikian metamorphism). Detailed studies may reveal additional subgreenschist facies rocks within areas outlined as greenschist facies on the metamorphic map, and may enable more precise dating of this event.

Greenschist Facies

Greenschist facies rocks occur in three regions: a thin, narrow discontinuous zone along the north side of the Central Borden Fault Zone in the Mary River area; a narrow belt near Grant-Suttie Bay; and a larger area in the Central Piling Basin. The facies represents prograde Aphebian metamorphism, but locally merges with subgreenschist facies rocks discussed above that could have resulted from Helikian regional metamorphism.

Presence of chlorite, actinolite, clinozoisite, albite rather than oligoclase, and absence of staurolite, andalusite and cordierite are considered diagnostic. As data points are commonly several kilometres apart the greenschist-amphibolite facies boundary is located approximately, and is based on different criteria. Whereas plagioclase composition has been used regionally, the distribution of staurolite, cordierite, and alumino-silicate minerals was employed locally. Prehnite seems to belong to some equilibrium assemblages in greenschist facies terrane.

Some typical assemblages are:

Quartz-albite-microcline-muscovite ± (biotite, or biotite + chlorite)

Quartz-albite-muscovite-biotite-chlorite ± (calcite, or tourmaline)

Quartz-albite-chlorite-epidote ± magnetite

Quartz-albite-(actinolite-hornblende) + (garnet-magnetite, or chlorite-epidote-calcite)

Quartz-plagioclase-biotite-muscovite-clinozoisite-prehnite

Quartz-biotite-chlorite-prehnite + (actinolite-calcite, or oligoclase-andesine)

Quartz-andesine?-hornblende-clinozoisite

Quartz-muscovite-biotite-chlorite + (chloritoid, or tourmaline)

Quartz-muscovite-biotite-garnet-actinolite or hornblende-calcite + (chlorite, or clinozoisite)

Quartz-actinolite-chlorite-talc-serpentine

Quartz-chlorite-tremolite or actinolite-parankerite-calcite-magnetite

Actinolite or hornblende-chlorite-epidote + (muscovite-microcline, or diopside-talc)

Chlorite-garnet-biotite muscovite

Serpentine-chlorite, with one or more of actinolite, talc, cummingtonite

Accessory minerals include magnetite, sphene, apatite, and zircon. Tourmaline, common in Piling Group rocks, is rare in the Mary River Group.

Although Piling Group strata immediately southwest of Barnes Ice Cap are shown as undivided amphibolite facies on Map 1475A, greenschist facies assemblages occur locally. A typical assemblage is:

Diopside-tremolite-clinozoisite-prehnite-sphene-quartz

Other smaller areas of greenschist facies are present in southern Baffin Island west of Amadjuak Lake.

Lower Amphibolite Facies

Although rocks of this facies are more common than greenschist facies assemblages, they nevertheless underlie a very small part of Baffin Island. In the Mary River Group they occur as narrow lenses within higher grade terrane, and as rims around greenschist facies strata.

The largest area of lower amphibolite facies surrounds the Central Piling Basin greenschist zone. Much of the undivided amphibolite facies in this region includes lower amphibolite facies rocks that cannot be differentiated with the available reconnaissance data.

Several small lenses also occur in the Hoare Bay Group on Cumberland Peninsula, and in supracrustal rocks on eastern Hall Peninsula. The lower amphibolite facies is rare south of Nettilling Lake and Cumberland Sound, but is associated with scattered greenschist facies areas west of Amadjuak Lake.

Presence of oligoclase or more calcic plagioclase with or without hornblende, staurolite, cordierite with biotite, and andalusite, and absence of chlorite in contact with muscovite have been used to identify lower amphibolite facies assemblages. Muscovite seems to co-exist with hornblende in the same assemblage at a few localities.

The metamorphosed supracrustal rocks contain a great variety of mineral assemblages. The more interesting and diagnostic associations from Mary River Group pelitic rocks at this grade are listed below. Mineral associations for metamorphosed iron formation, mafic and ultramafic igneous rocks in the Mary River Group, although chiefly from the lower amphibolite facies, also include assemblages from greenschist, upper amphibolite and granulite facies areas. These are listed without attempting to define their grade. The mafic and ultramafic rocks are commonly highly retrogressed.

Mary River Group Pelite and Greywacke

Quartz-plagioclase-potash feldspar-biotite-muscovite-sillimanite ± (cordierite, or garnet)

Quartz-plagioclase-potash feldspar-biotite-muscovite-cordierite-chlorite

Quartz-plagioclase-biotite-hornblende + (diopside, or potash feldspar, or muscovite? and epidote)

Quartz-plagioclase-biotite-muscovite + (sillimanite, or calcite)

Quartz-plagioclase-garnet-cordierite-staurolite-sillimanite

Quartz-beryl-muscovite-andalusite

Quartz-biotite-garnet-magnetite + (muscovite, or muscovite and chlorite, or staurolite, or cordierite)

Quartz-anthophyllite-biotite-cordierite + (sillimanite, or plagioclase and magnetite)

Quartz-biotite-muscovite-cordierite-andalusite + (sillimanite and plagioclase, or kyanite, or sillimanite)

Quartz-biotite-garnet-cordierite-andalusite + (staurolite, or sillimanite and magnetite)

Quartz-biotite-cordierite-sillimanite + (garnet and staurolite, or hornblende)

Quartz-biotite-garnet-(hornblende-actinolite) + (calcite and chlorite, or muscovite?)

Garnet-staurolite

Garnet-chlorite-anthophyllite

In general, kyanite and staurolite are relatively rare in the Mary River Group, but sillimanite commonly occurs with andalusite or cordierite.

Two thin sections of finely foliated, porphyroblastic metapelite from one locality in the Mary River area contain assemblages indicative of pressure and temperature fluctuations close to the alumino-silicate triple point (6 kb, 600°C).

(1) Biotite-cordierite-kyanite-andalusite-staurolite-sillimanite-quartz

(2) Biotite-andalusite-sillimanite-muscovite-quartz

In assemblage (1) biotite-cordierite-kyanite-staurolite-quartz may have preceded the andalusite, which in turn may have preceded the sillimanite. In assemblage (2) the sillimanite may also be later and is fibrolite formed at the expense of biotite.

Mary River Group Iron Formation

Quartz-cummingtonite ± magnetite ± hornblende

Quartz-hornblende-magnetite-diopside-carbonate

Quartz-cummingtonite-garnet ± chlorite

Quartz-cummingtonite-diopside ± carbonate

Quartz-(tremolite-actinolite)-diopside-magnetite

Cummingtonite-magnetite + (plagioclase, or garnet and biotite, or garnet and chlorite)

Quartz-anthophyllite-cordierite-biotite-magnetite ± garnet

Quartz-cordierite-muscovite-sillimanite-magnetite

Quartz-oligoclase-biotite-muscovite-sillimanite-cordierite-magnetite

Parankerite-quartz-chlorite-calcite-(tremolite-actinolite)-magnetite

Quartz-anthophyllite-magnetite ± (plagioclase and biotite)

Quartz-biotite-muscovite-garnet-magnetite ± chlorite

Albite-epidote-magnetite ± chlorite

Quartz-actinolite-pyrite-pyrrhotite-magnetite

Anthophyllite-cordierite ± quartz ± magnetite

Mary River Group Basalt, Gabbro, and Anorthosite

Plagioclase-hornblende-diopside-epidote + (microcline and calcite, or quartz and talc and calcite)

Plagioclase-hornblende-diopside-biotite-chlorite ± quartz

Plagioclase-hornblende-diopside-clinozoisite-garnet

Plagioclase-hornblende-anthophyllite-biotite-muscovite

Plagioclase-hornblende-cummingtonite ± (biotite and quartz and magnetite)

Hornblende-biotite-quartz-clinozoisite

Chlorite-biotite-muscovite ± (garnet and staurolite)

Anthophyllite-chlorite-quartz

Plagioclase-hornblende-epidote

Hornblende-epidote-diopside + (talc ± garnet, or potash feldspar, or garnet)

Mary River Group Ultramafic Rocks

Many assemblages contain relict primary minerals which are not given:

Serpentine-actinolite-talc-magnetite + (chlorite or calcite)

Serpentine-actinolite + (magnetite or chromite)

Serpentine-cummingtonite-chlorite-talc-magnetite

Hornblende-chlorite-magnetite

Garnet-anthophyllite-chlorite

Piling Group

Lower amphibolite facies assemblages from pelitic rocks in this group include:

Quartz-plagioclase-biotite-muscovite-cordierite-sillimanite ± (potash feldspar and garnet)

Quartz-plagioclase-microcline-muscovite?-diopside-epidote + (tremolite and scapolite, or hornblende and prehnite?)

Quartz-microcline-biotite-muscovite

Quartz-plagioclase-potash feldspar-biotite-muscovite ± (tourmaline or sillimanite)

Quartz-plagioclase-potash feldspar-muscovite ± (epidote, or phlogopite and sillimanite)

Quartz-plagioclase-clinopyroxene-hornblende-epidote

Quartz-biotite-garnet-anthophyllite-cummingtonite-cordierite-sillimanite

Quartz-plagioclase-biotite-chlorite-epidote

Quartz-plagioclase-biotite-muscovite + (tourmaline, or sillimanite and tourmaline, sillimanite and garnet ± tourmaline, or garnet and anthophyllite or cordierite and staurolite, or andalusite, or garnet)

Quartz-plagioclase-biotite-muscovite-kyanite

Quartz-microcline-garnet-muscovite-sillimanite

Kyanite and staurolite are as rare in the Piling Group as in the Mary River Group. Tourmaline occurs sporadically and seems more common along the northern margin of the Piling Basin. It occurs in uppermost lower amphibolite facies rocks with an affinity for muscovite- and muscovite-sillimanite-bearing schists.

Assemblages from iron formation, mafic and ultramafic rocks in the Piling Group are similar to those recorded from the Mary River Group, and include:

Quartz-magnetite + (actinolite or cummingtonite)

Quartz-plagioclase-hornblende-cummingtonite-magnetite ± (biotite and garnet)

Actinolite-biotite-magnetite

Actinolite-plagioclase + (quartz, or plagioclase and anthophyllite)

Actinolite-phlogopite-epidote-chlorite-magnetite

Hornblende-garnet-quartz-biotite ± (plagioclase and orthoclase)

Hornblende-biotite-magnetite + (scapolite and/or quartz and plagioclase)

Hornblende-actinolite-diopside-talc-microcline-calcite

Hornblende-biotite-calcite

Actinolite-chlorite + (biotite or phlogopite)

Tremolite-phlogopite-magnetite

Some typical associations from carbonate rocks are:

Carbonate-phlogopite-(tremolite-actinolite)-scapolite

Carbonate-phlogopite-olivine-garnet-amphibole-diopside

Carbonate-(tremolite-actinolite)-diopside-plagioclase-microcline-scapolite ± (quartz and talc? and epidote)

Carbonate-(tremolite-actinolite)-diopside-microcline-scapolite

Grossularite-diopside-quartz ± carbonate

Quartz-microcline-phlogopite-scapolite-calcite

Quartz-hornblende-zoisite

Quartz-biotite-hornblende-diopside

Calcite-talc-serpentine-forsterite-diopside

Calcite-tremolite + (phlogopite or diopside)

Hoare Bay Group

The following assemblages have been recorded from this group in southern Cumberland Peninsula:

Quartz-plagioclase-biotite-muscovite-sillimanite + (andalusite, or cordierite, or myrmekite, or garnet and tourmaline)

Quartz-plagioclase-biotite-cordierite-sillimanite-andalusite-staurolite ± (garnet or tourmaline)

Quartz-plagioclase-microcline-biotite ± (tremolite-actinolite)

Quartz-biotite-sillimanite-staurolite-andalusite

Kyanite, and to a lesser extent staurolite, are rare in Hoare Bay metapelites. Tourmaline appears to be more abundant in the Hoare Bay Group than in the Mary River Group but is not as common as in the Piling Group.

Hall Peninsula Strata

Assemblages that have been reported from supracrustal sequences on eastern Hall Peninsula include:

Quartz-plagioclase-biotite-muscovite-sillimanite ± garnet

Quartz-plagioclase-muscovite-garnet

Quartz-plagioclase-microcline-biotite-muscovite-sillimanite

Quartz-plagioclase-biotite-muscovite-tourmaline-garnet

Cordierite-biotite-muscovite-sillimanite

Biotite-muscovite-kyanite-sillimanite-andalusite

One alumino-silicate triple point assemblage was found on Hall Peninsula where lower amphibolite facies metamorphism was presumably similar to that in the Mary River area. Cordierite appears to be relatively rare on Hall Peninsula.

Upper Amphibolite Facies

Aphebian upper amphibolite facies rocks underlie the bulk of two large regions on Baffin Island; one extending north from Barnes Ice Cap, the other including southeast Cumberland Peninsula and east Hall Peninsula. A small, irregular belt also bounds the south side of the Piling Basin, and several small areas south and west of Amadjuak Lake contain minor greenschist and lower amphibolite facies rocks.

The upper amphibolite facies was located by the absence of staurolite and also by the absence of muscovite in the presence of quartz and plagioclase, and by the occurrence of potash feldspar, alumino-silicate minerals, cordierite, and garnet. Potash feldspar is chiefly microcline; and myrmekite, perthite, antiperthite and accessory rutile are common. In the absence of other criteria anatexis was taken to indicate upper amphibolite conditions.

Mineral assemblages from supracrustal sequences in seven regions, shown in Figure 1, are listed in Table 3. Several of these also occur with rocks of higher or lower grade. Whereas accessory opaque minerals, chiefly magnetite, are ubiquitous, tourmaline and epidote are of sporadic occurrence. Scapolite, kyanite, and andalusite are rare. Cordierite is less common in regions 3 and 4, and is relatively rare along the northeastern edges of regions 5 and 6 in both lower and upper amphibolite facies rocks. Phlogopite is characteristic of carbonate and of mafic and ultramafic rocks, which also contain cummingtonite and anthophyllite.

The metamorphic grade of the foliated and gneissic granitoid rocks is uncertain. Presence of deep red-brown biotite, green-brown hornblende, rutile, myrmekite, perthite, antiperthite and abundant anatexis is therefore taken to indicate attainment of at least upper amphibolite facies.

Granulite Facies

Granulite facies rocks occupy most of southern Baffin Island and underlie a complex region extending north from Meta Incognita Peninsula to north of Nettilling Lake and the Penny Ice Cap. A northeast-trending belt, cutting the regional strike, crosses Baffin Island north of Barnes Ice Cap. East Bylot Island is chiefly granulite facies and several small areas also occur in northeast Baffin Island.

The occurrence of metamorphic orthopyroxene, chiefly hypersthene, was taken to indicate the lower boundary of the granulite facies. Primary relicts of orthopyroxenes are present in meta-igneous mafic and ultramafic rocks at some localities and may have been mistaken for their metamorphic counterparts.

Quartz-clinopyroxene-garnet ± hornblende and/or biotite, plagioclase and potash feldspar assemblages commonly occur with hypersthene-bearing rocks, and have therefore been used in some cases to indicate granulite metamorphism. Preliminary microprobe analyses and refractive index – specific gravity – d spacing determinations indicate that garnets in the granulite and upper amphibolite facies are chiefly almandine with a large proportion of the pyrope end member. Assemblages for seven supracrustal regions are listed in Table 4. Some of the assemblages listed may occur in lower metamorphic grades. Myrmekite, perthite, antiperthite and accessory rutile are more common than in upper amphibolite facies assemblages. In the granulites orthoclase is at least as common as microcline. Garnet is common except in region 3. Sillimanite, cordierite, cummingtonite and anthophyllite are uncommon in all seven regions. Cummingtonite is most abundant in region 3, where it occurs with enstatite and hypersthene. Phlogopite and chromite are present chiefly in region 3. A corundum-bearing rock on southern Bylot Island (region 7) contains no hypersthene although this is present in contiguous outcrops. Sapphirine has been reported in granulites near Lake Harbour (region 1; W.L. Davison, pers. comm.).

Clinopyroxene-garnet-quartz assemblages, absent only in region 3, indicate relatively high pressure granulite facies metamorphism and also occur in the small granulite area on eastern Borden Peninsula (Figs. 1, 3). The only clinopyroxene-garnet-bearing assemblage that is not associated with granulite facies rocks is in region 3 on the southeast coast of Cumberland Peninsula and this rock contains abundant hornblende. Clinopyroxene-garnet occurs without quartz, hypersthene or plagioclase, east of Nettilling Lake in region 4 but is in a marble and so is probably not diagnostic of high pressure. Nearly all of the other clinopyroxene-garnet-quartz-bearing assemblages contain both plagioclase and hornblende, and most also contain hypersthene. Where plagioclase is absent, hornblende, hypersthene, and/or biotite, are present. Therefore, most of these assemblages represent the transitional stage to the relatively high pressure clinopyroxene-almandine-quartz granulite metamorphism. High pressure granulite metamorphism may be represented by the plagioclase- and/or hypersthene-bearing assemblages in areas 1, 4 and 6, which also contain small amounts of biotite but no hornblende. Rare occurrences of sillimanite, sapphirine and kyanite also indicate medium to high pressures.

More than one retrogressive recrystallization may have affected the granulites. Some granulite and monzocharnockite equilibrium mineral assemblages in the upper Cumberland Sound region contain brown hornblende partially recrystallized to secondary blue-green hornblende which in turn is partly altered to actinolite at some localities. In addition, a slight, very low grade retrogressive recrystallization has affected the granulite and amphibolite facies rocks regionally.

Granulites within the three major granulite regions (Fig. 3) display different boundary relationships with adjacent monzocharnockitic rocks, which occur chiefly in the core areas of all three regions, and with lower grade terrane. This is due in part to differences in the isograd surfaces and their orientations. On Bylot Island the upper amphibolite-granulite transition occurs over a short distance, and the hypersthene isograd surface is broadly conformable with the regional structure. The southwestern part of this granulite zone, and

Table 3

Upper amphibolite facies co-existing mineral assemblages for supracrustal strata

Mineral	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37	38
Quartz	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+
Plagioclase	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+					+	+	+
Potash-feldspar	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+													+	+	+	+			
Biotite	+	+	+	+	+	+	+	+	+	+	+	+	+	+						+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	7			
Phlogopite																		5																				
Anthophyllite																															+							
Cummingtonite																			+					±3		±5												
Tremolite																		5																				3
Actinolite															5													+	+							5		
Hornblende	+	+	+																	+	+			3		+		5				5			+	+	+	
Clinopyroxene		+								+					+					+				+													+	+
Garnet				+		+	+	+	+			+				+	+					+		3	5	+	+	+			3	+	+	+		5		
Andalusite																														+								
Cordierite	1				+		+	+								+					4		+				±3			+	3	4		+			4	
Sillimanite				+	+	+	+				5		+			+	+	5					6		+		+			+			+	+				
Scapolite															4																							
Carbonate															4							±3																+
Epidote			+						5					+									4						+							3		
Tourmaline																					1						±3			+								
Spinel						+						+	+					1			6				5							±4	+					
Magnetite																						±3																
Rutile				1																										+								
Olivine																																5		5				
Talc																																						
Wollastonite																																						
Chromite																																						
Region																																						
1	*	*		*	*	*				*	*				*	*	*			*	*	*		*									*		*	*		
2	*	*	*	*			*		*		*			*			*			*	*		*															
3	*			*				*	*	*	*	*		*						*	*	*	*		*	*	*	*	*	*							*	
4	*			*				*	*		*		*																			*		*				
5	*	*	*		*	*	*	*	*			*				*				*	*	*				*						*				*		*
6	*		*	*	*		*	*	*									*	*		*	*									*							
7	*																				*											*						

Mineral	39	40	41	42	43	44	45	46	47	48	49	50	51	52	53	54	55	56	57	58	59	60	61	62	63	64	65	66	67	68	69	70	71	72	73	74	75
Quartz	+	+	+	+	+	+	+	+	+	+	+	+																									
Plagioclase	+	+											+	+	+	+	+	+	+	+																	
Potash Feldspar													+																								
Biotite			+	+	+	+	+	+		5			+	+	+	+					+	+	+														
Phlogopite									+								+							+	+	+	+	+	+								
Anthephyllite							+										+									+											
Cummingtonite	+														6									+	+	+		+									+
Tremolite					+													+									+			+							
Actinolite																					+			±							+	+	+				
Hornblende	+	+				5	+		+	5	+	+			+	+			+	+		+	+						+		+						
Clinopyroxene		+				+			+		+		+			+	+	+	1	+		+	+					+		+	+			+	+	+	
Garnet	+	+	+	+								+		+									+												+		
Andalusite																																					
Cordierite				+																																	
Sillimanite			6	3				+																									+			+	
Scapolite																																					
Carbonate	+				+				+		5				+		+					4										+	+	+	+	+	
Epidote																								+													
Tourmaline			1																																		
Spinel			6																	1	+	4				+				+				+			+
Magnetite			6																																		
Rutile																																					
Olivine										+							+			1	+	+			+	+			+						+		+
Talc																																+					
Wollastonite																																				+	
Chromite																																					+
Region																																					
1			*		*		*					*	*	*	*	*				*	*																
2		*																*		*			*						*								
3	*																*					*		*	*	*		*									*
4								*																											*	*	
5						*			*																		*			*	*	*	*				
6				*			*			*					*				*															*			
7											*					*																					

regional numbers (1 to 7) refer to Figure 1;
+ primary assemblage in individual thin section;
* region the assemblage occurs in;

± mineral not in all assemblages;
±3 mineral not in all assemblages in region 3;

regional numbers (1 to 7) mineral(s) that occur with the primary assemblage in an additional assemblage in the region indicated

Table 4

Granulite facies co-existing mineral assemblages for supracrustal strata. Explanation as for Table 3

Mineral	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22
Quartz	+	+	+	+	+	+	+	+	+	+	+	±	+	+	+	+	+	+	+	±	4	+
Plagioclase	+	+	+	+	+	+	+	+	+	+	+	+			+	+	+	+	5	±	±4	
Potash Feldspar	+	+	+	4	±4	+	+						+	1								
Biotite	+	+	+	+	±			+	+	+	+	±	+	+								+
Phlogopite																						
Anthophyllite																						
Cummingtonite										5												
Tremolite-Actinolite																						
Actinolite																						
Hornblende			±	4	+			3	+			±	7			+	+	+		±		+
Clinopyroxene	1		+		+	1	+	+			+	±	7		+	+	+				+	1
Hypersthene	+	+	+	+	+	+	±	+	+	+		+	+	+	+	+		+	+	+		+
Bronzite																						
Enstatite																						
Garnet		+					+	6			+	±	7			+	+	+	+		+	+
Cordierite		±		±										+								
Sillimanite														1								
Spinel																				±		
Olivine																				+	+	
Corundum																						
Chromite																						
Talc																						
Carbonate																					+	
Magnetite																6			+			
Sapphirine																						
Region																						
1	*			*	*	*			*			*							*	*		
2			*						*								*					
3			*						*													
4	*	*	*		*		*	*	*	*	*	*	*		*						*	
5	*	*						*	*	*	*			*	*	*		*	*			
6	*				*			*	*							*			*			
7																	*					*

Mineral	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37	38	39	40	41	42	43	44	45
Quartz	+	±																					
Plagioclase			+	±	±	+	+	+	7	+	+												
Potash Feldspar			+	+																			
Biotite			+			+	+	±	+	+		+	+	+									
Phlogopite				+	+										+	+	3	+					
Anthophyllite															+								
Cummingtonite										±					+	+	3		4				
Tremolite-Actinolite																				+			
Actinolite														+									
Hornblende		+				+		+	+		+	+	+								+	+	
Clinopyroxene						±		+	+		+		+		±	+	3	3	+				
Hypersthene	+	+				+	+	+	7		+	+	+	+		+			+	+	+	+	+
Bronzite															+			+					
Enstatite																	+						
Garnet			+				+		+	+													+
Cordierite										+													
Sillimanite										+													
Spinel		+	+		+								+		±		3	+	+	+		+	
Olivine		+										+		+	+	+	+	+	+	1	+		
Corundum			+																				
Chromite																	3	3					
Talc											+										+		
Carbonate																+					+	+	
Magnetite	+																						
Sapphirine				+	+																		
Region																							
1		*		*	*			*				*	*						*		*		*
2									*					*						*			
3															*	*		*					
4																			*			*	
5							*			*													
6	*					*					*												
7			*														*						

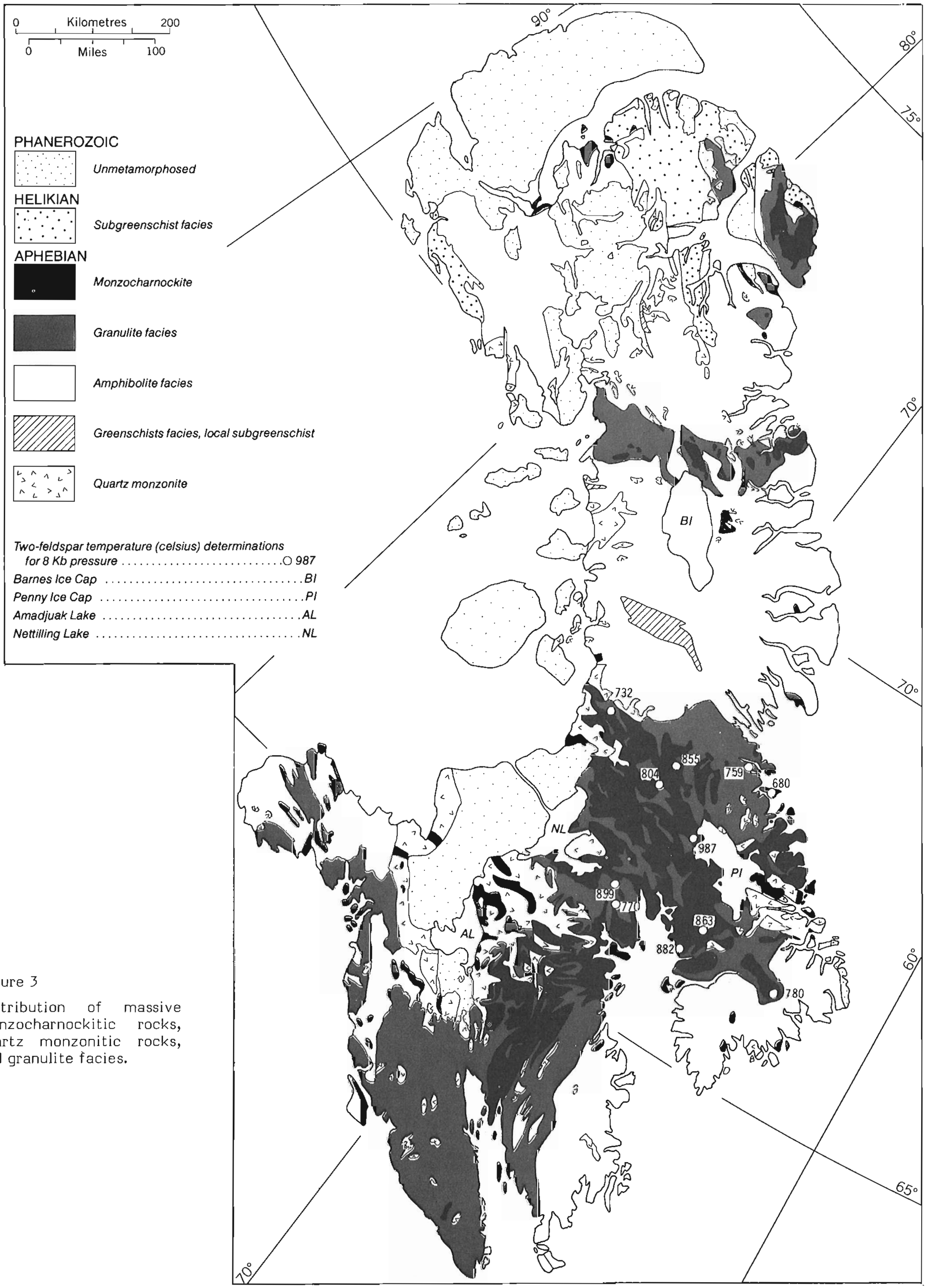


Figure 3

Distribution of massive monzocharnockitic rocks, quartz monzonitic rocks, and granulite facies.

the adjacent upper amphibolite facies, are sheared and contain closed recumbent folds overturned to the southwest (Jackson and Davidson, 1975).

The northeast-trending granulite belt at the north end of Barnes Ice Cap has more irregular boundaries. A relatively straight, abrupt part of the southern boundary northeast of Grant-Suttie Bay coincides with a fault zone. Elsewhere, this belt is bordered by an irregular, diffuse, transitional zone that is not related to retrogression.

The hypersthene isograd bounding the Barnes Ice Cap granulite belt generally transgresses the regional trend delineated by various map units but is also parallel to local structures, which may indicate deformation of the isograd surfaces.

The Barnes Ice Cap granulite belt is disrupted by northwest-trending prograde upper amphibolite zones that are gradational with granulite, and is cored in the northeast by charnockitic masses. This granulite belt occurs along a zone that separates northeast structural trends to the south from west and northwest trends to the north, and is on strike with a seismically active zone in Baffin Bay (Basham et al., 1977). This belt could have formed by a series of thermal domes aligned along a major structural break which has since been reactivated.

Boundaries of the vast granulite zone in southern Baffin Island are extremely complex. The hypersthene isograd is broadly conformable with structures and rock types in some areas but is transgressive elsewhere. Such complex relationships have been mapped in the Piling, Hoare Bay and Lake Harbour groups, in the large quartz monzonite-monzocharnockite bodies, and in most other rock types. North of Nettilling Lake and Cumberland Sound, and also west of Cumberland Sound, the hypersthene isograd is roughly parallel to the contact of the monzocharnockitic rocks that form the core of this region. In several places, such as north of Nettilling Lake, granulite-facies lenses commonly up to 1.6 km long and rarely 5 km long, occur within upper amphibolite facies rocks adjacent to the granulite boundary. Upper amphibolite lenses of similar size occur in the nearby granulites.

The Meta Incognita-Foxe Peninsula region is particularly complex, containing chiefly granulite and upper amphibolite facies rocks associated with only minor small quartz monzonite-monzocharnockite intrusions. Granulites are abundant in a broad zone adjacent to the southwest margin of the large massive quartz monzonite body southwest of Amadjuak Lake (Fig. 3).

MASSIVE QUARTZ MONZONITIC AND MONZOCHARNOCKITIC PLUTONS

The writers consider that the bulk of the monzocharnockitic rocks on Baffin Island are metamorphosed intrusions (Fig. 3). As such they are included within the granulite facies metamorphic terrane on Map 1475A and are not shown separately like the massive quartz monzonitic rocks.

The Geological Survey of Canada classification for igneous granitic rocks is used in this paper (Brown, 1952). Nomenclature for charnockitic, hypersthene-bearing equivalents employed here are: charnockite (granite), monzocharnockite (quartz monzonite or adamellite), and granoenderbite (granodiorite). Minor more mafic phases of both charnockitic and granitic rocks that are present include enderbite (quartz diorite), jotunite (syenodiorite) and anorthosite.

General Features

Massive quartz monzonite and the much more abundant monzocharnockite occur chiefly in south Baffin Island, where the two are closely related in space. Abundance of these suites is about equal in north-central Baffin Island, but there they are not spatially associated (Fig. 3). Both suites are generally in, or adjacent to, regions of granulite facies metamorphism.

The massive quartz monzonitic and monzocharnockitic bodies range in dimensions from dykes, sills, and irregular masses to a huge group of anastomosing bodies that occupies tens of thousands of square kilometres in southern Baffin Island (Fig. 3). Although most of these are not deformed or recrystallized, a higher proportion of the monzocharnockitic rocks are granoblastic. Most of the larger bodies are, however, elongated parallel to regional structures, suggesting that they have been involved in deformation. Both cross-cutting intrusive, and sharp concordant contacts occur with massive and layered rocks, including supracrustal strata, which also occur as inclusions. Many of the metasedimentary lenses within the south Baffin mass are synformal. In several places monzocharnockite, and to a lesser extent quartz monzonite, are intruded by younger granitic bodies that range up to 300 m thick sheets.

The massive quartz monzonitic and monzocharnockitic rocks are texturally and chemically similar, but the monzocharnockitic varieties are deficient in hydrous minerals. Quartz monzonite and monzocharnockite are most common, but there is also considerable granodiorite and granoenderbite with less granite and charnockite. Both rock suites are medium to coarse grained or pegmatitic, and include varieties with feldspar phenocrysts. Fresh quartz monzonitic rocks are chiefly pink or red, whereas monzocharnockitic varieties have a resinous lustre and are honey brown to greenish grey.

The most common assemblage in the quartz monzonitic rocks is quartz-plagioclase-microcline-biotite; other assemblages are listed in Table 5. Plagioclase tends to be antiperthitic and composition ranges from albite to andesine. Potash feldspar is commonly perthitic and orthoclase occurs locally. Myrmekite is common and muscovite is present chiefly in north-central Baffin Island. Hornblende, clinopyroxene, and garnet occur near granulite facies terrane or near monzocharnockitic rocks. Cordierite and sillimanite have been reported from localities near metasedimentary inclusions.

Monzocharnockitic rocks are characterized by the assemblage quartz-plagioclase-potash feldspar-biotite-hypersthene ± hornblende; other assemblages are listed in Table 5. Plagioclase is antiperthitic and ranges from oligoclase to labradorite. Potash feldspar tends to be perthitic and both microcline and orthoclase are abundant. Myrmekite, clinopyroxene, garnet and hornblende are common, and quartz-clinopyroxene-garnet assemblages indicate relatively high pressure, as with the granulites.

Age

Contact relationships indicate that these plutonic rocks are younger than the Archean and Aphebian supracrustal sequences. K-Ar mica ages from the quartz monzonitic and monzocharnockitic rocks range from about 1500 to 1800 Ma (Wanless, 1970). The U-Pb zircon age of 1900 ± 20 Ma from charnockitic granulite near Pangnirtung is considered by Pidgeon and Howie (1975) to date the igneous activity that preceded granulite facies metamorphism. This interpretation is in agreement with a Rb-Sr isochron age of 1670 ± 30 Ma obtained by the Geochronological Laboratories of the Geological Survey of Canada for charnockitic granite that intrudes granulite facies Piling strata about 80 km northwest of the northwest end of Penny Ice Cap, and with the 1812 ± 30 Ma Rb-Sr isochron at Frobisher Bay. The age determinations provide a minimum age for the massive quartz monzonitic and monzocharnockitic rocks, and indicate the age of the late Aphebian high grade metamorphism that affected these rocks, essentially obliterating traces of earlier events.

Origin

Archean gneissic and plutonic rocks on Baffin Island are compositionally similar to the Aphebian massive quartz monzonitic and monzocharnockitic rocks. The Aphebian plutons probably represent Archean crustal rocks melted at depth and intruded into higher crustal levels, although some bodies may have recrystallized in situ. The 1670 Ma Rb-Sr isochron for the charnockitic granite northwest of the Penny Ice Cap has a relatively high initial $^{87}Sr/^{86}Sr$ ratio of 0.7149 ± 0.0022, which supports the view that these intrusions have resulted from the anatexis of Archean rocks. The association of the massive plutonic rocks with high grade metamorphic terrane suggests that they are probably deep to intermediate level intrusions. Reconnaissance mapping indicates that the monzocharnockitic plutons are derived chiefly from granulite facies metamorphism of the massive quartz monzonitic intrusions but also include recrystallized Archean and Aphebian gneissic granitic rocks.

The Bylot Island Batholith is chiefly an intensely deformed feldsparphyric monzocharnockite (Jackson and Davidson, 1975) that locally contains a large proportion of nebulitic gneiss. The deformed monzocharnockite grades locally into undeformed massive to faintly foliated rock that can be interpreted as the result of later recrystallization. Elsewhere, the massive rock seems to be relict patches of undeformed but metamorphosed material. Monzocharnockitic dykes and sills intrude the granulite facies migmatite adjacent to the main intrusion. The monzocharnockite grades into less metamorphosed but deformed quartz monzonite along the north side of the pluton, which is a refolded sheet-like intrusion. Although the emplacement age is uncertain the batholith was metamorphosed in the late Aphebian.

The margin of the massive monzocharnockite body at the northeast end of the Barnes Ice Cap granulite belt is locally sheared and recrystallized. This body may contain primary as well as metamorphic hypersthene which at one locality is rimmed by diopside, which in turn is rimmed by hornblende. Some smaller monzocharnockitic bodies in this granulite belt extend beyond the hypersthene isograd into lower grade terrane where they are red, massive, hypersthene-free hornblende-biotite granitic rocks (Fig. 3).

The large anastomosing charnockitic mass in southern Baffin Island (Fig. 3) is composed of several rock types, but is chiefly the hypersthene-bearing equivalent of the quartz monzonitic rocks around its eastern and western margins. However, in the field, as granulite facies terrane is approached, it is increasingly difficult to differentiate between foliated granitic rocks, porphyroblastic migmatites and nebulitic gneisses. These rocks have a similar composition that ranges between quartz monzonite and granodiorite. In the granulite zone their textures, grain size, and mineral assemblages converge, structures are indistinct, and differences are masked by the highly weathered nature of the rocks. Also within the granulite facies it is very difficult to differentiate these rock types from metamorphosed massive quartz monzonitic rocks. The large monzocharnockitic mass in south Baffin Island may therefore include a sizeable proportion of these recrystallized older rocks.

Table 5
Co-existing mineral assemblages in massive quartz monzonitic and monzocharnockitic rocks

Mineral	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	A	B	C	D	E	F	G	H	I	J	K	L
Quartz	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+	+
Plagioclase	+	+	+	+	+	+	+	+	+	+	+	±	+	+	+	+		+	+	+	+	+	+	+	+	+	+	+	+	+	+
Potash Feldspar	+	+	+	+	+	+	+	+	+	+	+	+	+				+			+	+	+			+	+	+	+	+	+	
Biotite	+	+	+	+	+	+	+									+	+	+		+	±	±		+		+	+		+		±
Muscovite																					+						±				
Hornblende	±		+		±		+	±	+	+	±			±				+				+	+		±	+				+	
Garnet				+							+		+		+	±			+					±			+	+			
Clinopyroxene					+	+	+	+			+								+										+	+	
Hypersthene		+	+	±	+					+	+	+		+	+	+	+		+												
Region																															
Bylot Island		*	*	*						*	*								*												
Barnes Ice Cap	*		*		*	*	*		*	*					*			*													
S. Baffin Island	*	*	*	*	*	*		*			*	*	*	*		*	*														

Where massive quartz monzonitic and monzocharnockitic rocks are contiguous, no sharp intrusive contact is present. The chief differences are colour and presence or absence of hypersthene. The hypersthene isograd has been traced continuously between the quartz monzonitic and monzocharnockitic rocks, and through Archean gneiss, lit-par-lit and banded migmatites containing relicts of Piling, Hoare Bay and Lake Harbour groups, and through these metamorphosed supracrustal strata. This continuity of the hypersthene isograd is taken to indicate that hypersthene in the charnockitic rocks is chiefly metamorphic. Preliminary microprobe analyses of hypersthene show that their iron content is similar to that of the metamorphic rather than the igneous variety, as reported by Bhattacharya (1971).

Massive quartz monzonitic plutons are absent in granulite terrane throughout Baffin Island. Minor exceptions to this northeast of the Penny Ice Cap and on Meta Incognita Peninsula (Fig. 3) are chiefly in undivided upper amphibolite-granulite terrane.

Late Aphebian massive granitic and pegmatitic dykes, sills and sheets occur throughout most of the Aphebian and Archean terrane in concentrations that range from isolated occurrences up to 30 per cent or more of outcrops. Below granulite grade these minor intrusions are unmetamorphosed and have sharp contacts with adjacent rocks. In the granulite zone, however, they have a greasy lustre and diffuse contacts due to metamorphic recrystallization.

Most of the quartz monzonitic and monzocharnockitic plutons are considered to be sheet-like, or mushroom-shaped. The huge quartz monzonitic-monzocharnockitic mass in south Baffin Island is considered to be a series of coalescing intrusions and associated recrystallized basement areas. Another possibility is that this region represents the irregular upper surface of a widespread layer in the earth's sialic crust that was heated up and from which granitic magma rose as rootless intrusions that were emplaced only a short distance above. The emplacement environment may have ranged from anorogenic to post-orogenic, and from pre- to late metamorphic. The rarity of sillimanite, kyanite and sapphirine, and relative abundance of cordierite, suggest that the granulite-charnockite terrane was formed by relatively low pressure, high temperature metamorphism.

The assemblage clinopyroxene-garnet-quartz-hypersthene-plagioclase is stabilized at 700°C and at a pressure of 8-10 kb (Winkler, 1974). Charnockites with these minerals occur chiefly on Bylot Island, and in south Baffin Island. Within the granulites, however, these assemblages are present in 6 out of the 7 supracrustal regions. They are absent in region 3 and are most abundant in region 4. The potash feldspar-plagioclase geothermometer modified by Stormer (1975) indicated an average temperature of 838-860°C over 8-10 kb for 7 massive charnockitic samples, and an average temperature of 780-800°C over 8-10 kb for 5 metasedimentary granulite samples (Fig. 3). The highest temperature obtained was 1013°C at 10 kb for a monzocharnockitic sample from north of Cumberland Sound. The biotite-garnet geothermometer as described by Ferry and Spear (1977) indicated temperatures of 624-937°C at 0 kb and 667-989°C at 10 kb for three supracrustal granulite samples. These temperatures are somewhat higher than those provided by the geothermometer as modified by P. Nielsen (pers. comm.). Highest temperatures were obtained for a mineral pair whose garnet analysis indicates the presence of impurities. One sample indicated a temperature range of 624-662°C over 0-10 kb using the biotite-garnet geothermometer and 804-825°C over 8-10 kb using the feldspar geothermometer.

HELIKIAN METAMORPHISM

The Eqalulik and Uluksan groups of sedimentary and volcanic rocks on northwest Baffin and Bylot islands are tentatively considered to be of Neohelikian age (Jackson et al., 1975, 1978). One volcanic sample has given a whole-rock K-Ar age of 903 ± 140 Ma (Blackadar, 1970b). The Fury and Hecla, and Autridge formations on northern Melville Peninsula and adjacent Baffin Island (Blackadar, 1970b), and a small patch of sediments at Grant-Suttie Bay are correlated with the Eqalulik Group.

Secondary mineral assemblages within the Neohelikian Nauyat volcanics are, locally, more diagnostic of subgreenschist metamorphism than are secondary assemblages in the Hadrynian Franklin diabases. Three "anomalous" K-Ar ages of 963, 1005 and 1525 Ma obtained for Mary River Group rocks in the Mary River region could be related to the subgreenschist recrystallization in the region rather than to faulting and dyke emplacement, as previously suggested by Jackson (1978b). Because the age of the Neohelikian strata has not been well documented, their metamorphism is shown on Map 1475A as subgreenschist facies of a more general Helikian age.

The following metamorphic assemblages in Neohelikian strata are chiefly from the Nauyat volcanics:

Chlorite-albite-epidote-pumpellyite

Chlorite-muscovite-serpentine-calcite

Chlorite-hornblende-biotite

Chlorite-biotite-muscovite ± calcite

Chlorite-muscovite ± talc

Chlorite-prehnite-calcite

Biotite-muscovite ± albite

Albite-biotite + (calcite or muscovite)

Hornblende-albite-calcite + (biotite, or chlorite and serpentine)

HADRYNIAN ALTERATION

The Franklin diabase dykes have an age of about 675 Ma (Fahrig et al., 1971), and intrude all other Precambrian rocks on Baffin Island, but not lower Paleozoic strata. Throughout Baffin Island the extent of the secondary alteration present in the Franklin dykes ranges from greater to less than that present in the adjacent Archean-Neohelikian rocks. The secondary alteration of the Franklin diabases may be deuteric, although it is similar to the diaphthoresis present throughout the Archean, Aphebian, and Neohelikian rocks. Many of these secondary minerals could be the result of surface or near-surface alteration that took place chiefly on or adjacent to the present, lower Paleozoic, and Neohelikian erosion surfaces. The lower Paleozoic strata in the region, however, are not metamorphosed and contain very little secondary alteration material (Trettin, 1975). Mesozoic-Cenozoic strata are unmetamorphosed and most of the secondary minerals are inherited from the gneissic source terrane. Therefore the secondary alteration in the Franklin dykes is probably chiefly of Hadrynian age, whereas that in older rocks may contain Hadrynian as well as Helikian minerals.

Prehnite was identified locally in Hadrynian and older rocks, and has long been known to occur in quartz-calcite veins adjacent to Franklin dykes in Neohelikian strata on northwestern Borden Peninsula. Vesuvianite was reported from the same area (Bernier, 1912; Johnston, 1913). The following assemblage is from a Franklin dyke.

Chlorite-hornblende-epidote-prehnite + (serpentine, or biotite and carbonate)

EFFECTS OF POST-APHEBIAN DEFORMATION

Post-Aphebian events, chiefly of Helikian and Hadrynian age, have had a definite effect on Aphebian and Archean metamorphic rocks, which is discernible in the distribution of metamorphic facies and isotopic ages.

Aphebian K-Ar ages on Baffin Island range chiefly from 1500 to 1800 Ma (Wanless, 1970, pers. comm.; Wanless et al., 1974, 1978), and fall into two age groups that coincide with three northwest-trending major fault blocks. Most K-Ar ages north of Cumberland Sound-Nettilling Lake, and southwest of Frobisher Bay-Amadjuak Lake are at least 1655 Ma. Those on the intervening Hall Peninsula-Amadjuak Lake-Nettilling Lake block are 1605 Ma or less. Post-Aphebian northwest-trending vertical fault zones are common throughout Baffin Island, and movement has occurred on them from at least Helikian to Eocene time. Two such major fault zones are thought to separate the three blocks, in each of which topography suggests upward tilting towards the northeast. The Cumberland Peninsula block is the largest and extends northwest to Brodeur Peninsula, incorporating regions 3 to 7 on Figure 1. Franklin diabase dykes have been emplaced along the northwest-trending faults.

Metapelitic mineral assemblages for amphibolite facies rocks suggest different metamorphic pressures in the three major fault blocks. In the lower amphibolite facies, high pressure assemblages predominate on the Hall Peninsula block, whereas pressures are low on Cumberland Peninsula and moderate to high in regions 5 and 6 (Fig. 1). The distribution of upper amphibolite facies terrane implies that pressures were higher on Hall Peninsula than on the blocks to the north and south. Granulite facies assemblages indicate relatively low pressures for region 3 (Fig. 1). These conclusions suggest that the Hall Peninsula block has been uplifted more than the other blocks.

It seems likely that post-Aphebian folding and faulting have also modified the shape of the Barnes Ice Cap granulite belt. The northwest-trending upper amphibolite facies zones that disrupt this northeast-trending granulite belt are parallel to, and in some cases aligned with the following features: major open folds in Neohelikian strata, vertical faults that range from Neohelikian to Recent, and the Franklin diabase dykes. One of these northwest-trending zones is on strike with the major graben-syncline that contains Neohelikian strata and extends southeast from Borden Peninsula for 160 km.

Faulting has continued in the region to the present time. Neohelikian strata are more faulted than Paleozoic rocks, which are more faulted than Cretaceous-Eocene strata, and glacial lake beaches east of Steensby Inlet are only slightly offset. Concentrations of earthquake epicentres for the period 1962-1974 (Basham et al., 1977) coincide with the south and north edges of the northeastern part of Barnes Ice Cap granulite belt and extend northward from the northeast end of the belt. Epicentre concentrations also coincide with the northern edge of the south Baffin Island monzocharnockitic complex (Fig. 3), with northwest-trending crushed zones in quartz monzonitic and monzocharnockitic rocks northeast of the Penny Ice Cap, and with other geological and topographic features.

SUMMARY AND CONCLUSIONS

That part of Map 1475A covering Baffin and Bylot islands is based chiefly on results of four Geological Survey of Canada helicopter reconnaissance mapping projects supplemented by studies of smaller areas.

A large part of Baffin Island is underlain by Archean granitic rocks, emplaced or metamorphosed about 2800 Ma ago. The supracrustal Mary River Group was deposited about 2750 Ma ago. Kenoran orogenesis, granodioritic intrusion and metamorphism affected most of the Archean rocks 2550-2600 Ma ago. By early Aphebian time, most of the Archean terrane had been metamorphosed to amphibolite facies. Although the Mary River Group was intruded by granitic rocks, locally, the group remained unmetamorphosed or was metamorphosed only slightly.

Following epeirogenic uplift and erosion three extensive lower Aphebian supracrustal sequences, the Piling, Hoare Bay and Lake Harbour groups, were deposited in shelf, miogeoclinal, and eugeoclinal environments. There is meagre evidence locally of orogenesis in the 2100-2300 Ma time span.

Anorogenic to synorogenic granitic plutons were emplaced about 2000-1900 Ma ago and intense Hudsonian orogenesis and metamorphism affected the region 1964-1600 Ma ago. At present, it is not possible to subdivide this major event. Prograde subgreenschist to upper granulite facies mineral assemblages were formed and only small local areas were unaffected. In general the amphibolite facies metamorphism is a low pressure-high temperature type similar to Abukuma metamorphism, but relatively high pressures are indicated for Hall Peninsula, and possibly for east-central Baffin Island.

Granulite metamorphism continued in several places after deformation had largely ceased. Relatively high pressures are indicated locally at several widely scattered areas on Baffin and Bylot islands, and pressures were probably somewhat higher, relative to temperatures, than in the amphibolite facies terrane.

The massive quartz monzonitic and monzocharnockitic rocks represent low to intermediate level plutons, some of which may have melted and recrystallized in situ. The granulite facies event has resulted in metamorphic convergence of several rock types of similar composition, and the charnockitic rocks are thought to include a considerable amount of metamorphosed Archean and Aphebian granitic material. Preliminary pressure-temperature calculations with the two-feldspar geothermometer modified by Stormer (1975) indicate relatively low temperatures around the margins of the granulite-monzocharnockite complex north of Cumberland Sound and relatively high temperatures in the central part where a temperature of 1013°C is indicated for a pressure of 10 kb or 897°C for 8 kb. Preliminary determinations for biotite-garnet assemblages, as outlined by Ferry and Spear (1977), indicate lower temperatures. The granulite-monzocharnockite terrane of south-central Baffin Island probably represents the irregular upper part of an extensive layer within the sialic crust, with different ages in the layer reflecting differences in the time of uplift. The layer may represent material which had been kept hot for a long period of time and from which Aphebian deep to intermediate level plutons were derived.

As much as 20 km (12 miles) of uplift may have occurred following the Kenoran Orogeny, and up to 30 km (18 miles) of uplift – most of it prior to Neohelikian sedimentation and volcanism – may have occurred in parts of Baffin Island following the Hudsonian Orogeny. These two figures are local maxima and not necessarily complementary. The present Baffin Island crust is considered to be about 32 km thick (Jackson et al., 1977) but it was probably 60-70 km thick at some time before the Helikian.

Partial recrystallization ranging from diagenesis to subgreenschist metamorphism affected Neohelikian strata and possibly Hadrynian diabase dykes, but not Phanerozoic strata. The present distribution of metamorphic facies has been modified by gentle late Proterozoic southeast-trending folds and by southeast faulting that has been active from the Helikian to the present. The faulting has divided Baffin Island into three major blocks which have been tilted upward toward the northeast. The central block includes Hall Peninsula, and the presence there of relatively high-pressure amphibolite facies assemblages, together with somewhat younger late Aphebian K-Ar ages, indicates that this block has been uplifted relative to the regions to the south and north.

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METAMORPHISM OF THE MIDDLE PRECAMBRIAN (APHEBIAN) ROCKS OF THE EASTERN SOUTHERN PROVINCE

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Card, K.D., Metamorphism of the Middle Precambrian (Aphebian) rocks of the Eastern Southern Province; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 269-282, 1978.

Abstract

Middle Precambrian (Aphebian) supracrustal rocks and associated intrusions of the eastern Southern Province in Ontario were deformed and metamorphosed during several Middle Precambrian events. Deformation commenced prior to emplacement of Nipissing Diabase intrusions at 2150 Ma and deformation and a regional metamorphic culmination occurred at approximately 1900 Ma during the Penokean Orogeny. Finally, deformation and metamorphism followed emplacement of the Sudbury Nickel Irruptive at 1840 Ma. Mineral assemblages in the Middle Precambrian rocks indicate conditions ranging from diagenetic alteration to the amphibolite facies of regional metamorphism. The patterns of regional metamorphism are related to variations in the stratigraphy of the supracrustal rocks, to intensity and style of deformation, and to felsic plutonic activity. All are probably related to subcrustal processes localized along a major structural discontinuity in the Canadian Shield.

Résumé

Les roches supracrustales du Précambrien moyen (Aphébien) et les intrusions associées de l'est de la province du Sud en Ontario ont été formées et métamorphisées au cours de plusieurs évènements, qui ont eu lieu pendant le Précambrien moyen. La déformation a commencé avant la mise en place des intrusions de diabase de Nipissing il y a environ 2 150 Ma; pendant l'orogenèse du Pénokéen, qui date d'environ 1 900 Ma, a eu lieu une période de déformation et de culmination du métamorphisme régional. Finalement, une autre période de déformation et de métamorphisme, datant de 1840 Ma, a succédé à la mise en place de la masse intrusive nickelifère de Sudbury. Les assemblages minéraux des roches du Précambrien moyen indiquent que le métamorphisme régional varie entre une altération diagénétique et le faciès des amphibolites. Les traits du métamorphisme régional sont liés aux variations stratigraphiques des roches supracrustales, à l'intensité et au style de la déformation, et à l'activité plutonique felsique. De plus, ils sont tous probablement liés aux processus subcrustaux produits le long d'une discontinuité structurale importante du Bouclier précambrien.

INTRODUCTION

Middle Precambrian (Aphebian) rocks of the Southern Province in Ontario form a linear belt that extends along the north shore of Lake Huron, from Sault Ste. Marie to Sudbury, and northeastward to the Cobalt-Kirkland Lake area. These rocks, including strata of the Huronian Supergroup, gabbroic intrusions of the Nipissing Diabase suite, the Sudbury Nickel Irruptive and the Whitewater Group of the Sudbury Basin, were variably deformed and metamorphosed during the Middle Precambrian about 1600 to 2600 Ma ago. The Huronian sequence unconformably overlies Early Precambrian granitic and metavolcanic rocks of the Superior Province to the north, and along the Grenville Front Tectonic Zone (Lumbers, 1975) the zone of contact between the Grenville and other structural provinces of the Canadian Shield has been overprinted by the deformation and metamorphism characteristic of this zone. Southeast of the Grenville Front, Middle Precambrian rocks have been involved in the Late Precambrian high rank metamorphism and deformation characteristic of the Grenville Province. To the south, they are covered by undeformed Paleozoic strata (Fig. 1).

The eastern Southern Province can be subdivided into a number of areas that display differences in regional metamorphic grade, in thickness and facies of Middle Precambrian supracrustal rocks, in intensity and style of deformation, and in felsic plutonic activity. These subdivisions, shown in Figure 1, include the Sault Ste. Marie-Elliot Lake area, the Cobalt Embayment, the Sudbury Structure, and the Sudbury-Cutler area.

Regional metamorphism of the low pressure intermediate type (Miyashiro, 1961; Winkler, 1967) occurred under conditions ranging from diagenetic alteration to amphibolite facies. Although there is evidence for several metamorphic events, the higher grade assemblages are apparently in equilibrium and formed during one major metamorphic episode. The available geological and radiometric age data indicate that this metamorphic culmination occurred about 1900 Ma ago after emplacement of the Nipissing Diabase intrusions (2150 Ma) and prior to emplacement of the Sudbury Nickel Irruptive (1840 Ma) (Van Schmus, 1965; Fairbairn et al., 1969; Krogh and Davis, 1974). Recent studies (Van Schmus, 1976) indicate that major metamorphism and deformation occurred in the western part of the Southern Province approximately 1850 to 1900 Ma ago and consequently regional metamorphism and deformation occurred throughout the Southern Province at that time. There is also geological, radiometric, and paleomagnetic evidence for earlier, pre-Nipissing Diabase deformation and mild metamorphism approximately 2200 Ma ago, and for later thermal metamorphic and plutonic events approximately 1600 to 1750 Ma ago (Table 1).

METAMORPHIC MINERAL ASSEMBLAGES

The Middle Precambrian (Aphebian) rocks of the eastern part of the Southern Province were altered under conditions ranging from diagenesis-epigenesis, herein termed "subgreenschist facies" to the amphibolite facies of regional metamorphism. The rocks of the region can be subdivided, on the basis of the metamorphic mineral assemblages, into the following facies:

1) Subgreenschist facies
   – diagenetic and epigenetic zones

[1] Ontario Geological Survey, Ontario Ministry of Natural Resources

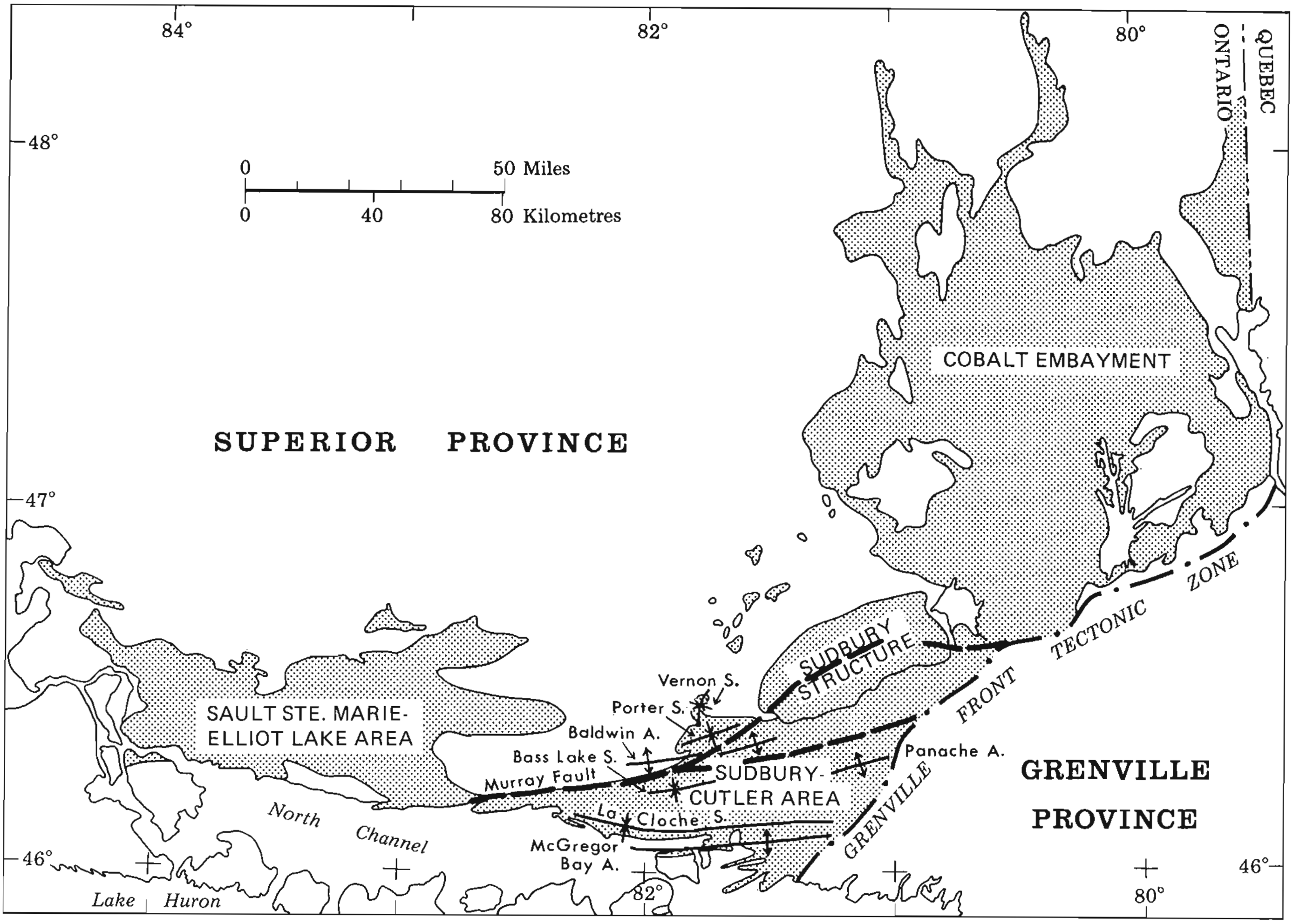

Figure 1. Location, subdivisions and major structural elements of the eastern Southern Province in Ontario.

2) Low to middle greenschist facies
   – chlorite and biotite zones
3) Middle to upper greenschist facies
   – garnet and chloritoid zones
4) Amphibolite facies
   – staurolite zone

In addition, adjacent to the Grenville Front felsic plutons there is a relatively narrow contact metamorphic aureole developed in the Huronian rocks with mineral assemblages corresponding to the hornblende hornfels facies (Card, 1976b).

For the purpose of describing their metamorphic mineral assemblages, the rocks can be subdivided into the following types:

1) Quartz and quartz-feldspar metasediments – primarily Huronian metasandstone, including aluminosilicate-bearing metasandstone.
2) Calcareous metasediments – metamorphosed limestone, dolostone, silty and sandy limestone and dolostone, and carbonate-bearing siltstone and sandstone, principally of the Huronian Espanola Formation.
3) Argillaceous metasediments – metamorphosed mudstone, shale, and wacke that occur throughout the Huronian sequence, notably in the McKim, Pecors, Espanola, Gowganda and Gordon Lake formations.
4) Mafic meta-igneous rocks – including mafic (basaltic) metavolcanics of the Huronian Supergroup and mafic (metagabbro) intrusions of the Nipissing Diabase suite.

Chemical analyses of rocks typical of the foregoing groups are given in Table 2.

Subgreenschist Facies

In the subgreenschist zone, all of the foregoing rock types consist of mixtures of original (detrital, chemically precipitated, or magmatic) minerals, diagenetic minerals, and epigenetic minerals. Assignment of particular minerals or mineral assemblages to one or other of the foregoing categories is difficult, especially the distinction between diagenetic and epigenetic minerals.

The following assemblages occur in the quartz and quartz-feldspar metasediments:

quartz – kaolinite

quartz – diaspore

quartz – pyrophyllite - sericite

quartz – hydromicas

quartz – muscovite (paragonite) - albite

Argillaceous rocks contain:

quartz – kaolinite

quartz — (illite - montmorillonite)

quartz — pyrophyllite - muscovite (paragonite)

quartz — muscovite (paragonite) - chlorite - albite

quartz — stilpnomelane

Calcareous rocks consist primarily of recrystallized calcite and dolomite with variable amounts of clay minerals, muscovite, chlorite, and talc. Mafic igneous rocks contain uralitic amphiboles, talc, epidote group minerals, magnetite and leucoxene. In addition, all of the foregoing rock types contain original sedimentary and igneous minerals. According to the scheme of regional epigenetic and metagenetic (subgreenschist) facies outlined by Kossovskaya and Shutov (1965), these assemblages would correspond for the most part to the "deep seated epigenesis" and "early and late metagenesis" stages.

Low to Middle Greenschist Facies

Rocks of the low to middle greenschist facies (chlorite and biotite zones) are partly recrystallized and still commonly contain original as well as metamorphic minerals. The following assemblages are present in quartzitic rocks:

muscovite — quartz

muscovite — andalusite - pyrophyllite - albite - quartz

muscovite — biotite - albite - quartz

Calcareous rocks contain:

calcite (dolomite) — muscovite - talc - chlorite

quartz — calcite (dolomite) - chlorite - biotite

scapolite — muscovite - epidote

scapolite — muscovite - biotite (phlogopite) - epidote

Table 1

Rock units and events

Event	Rock Unit	Radiometric Age (Ma)
PLEISTOCENE AND RECENT Glaciation	Till, glaciofluvial and glaciolacustrine deposits	0.1+
	UNCONFORMITY	
PALEOZOIC ORDOVICIAN-SILURIAN	Limestone, shale	
	UNCONFORMITY	
PRECAMBRIAN		
LATE PRECAMBRIAN (HELIKIAN)		
Grenville Orogeny		1000-1200
Mafic intrusion	Late Diabase Dykes	1280
Felsic intrusion	Croker Complex	1500
MIDDLE PRECAMBRIAN (APHEBIAN)		
Deformation, metamorphism (Hudsonian Orogeny?)	Grenville Front and Cutler Plutons	1600-1750
Sudbury Event	Nickel Irruptive Whitewater Group	1844
Deformation, metamorphism (Penokean Orogeny)		1900
Mafic intrusion	Nipissing Diabase	2150
Early deformation Felsic intrusion	Murray, Creighton Plutons	2200
Faulting, basin subsidence volcanism, sedimentation	Huronian Supergroup	
	UNCONFORMITY	
EARLY PRECAMBRIAN (ARCHEAN)		
Mafic intrusion Felsic intrusion, Deformation and metamorphism (Kenoran Orogeny)	Late mafic dykes Quartz monzonite plutons	2600

Table 2

Chemical analysis of typical metasedimentary and meta-igneous rocks of the Sudbury-Cutler Area

	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	55.07	62.00	60.73	59.30	49.97	48.34	52.4	51.2	61.48	4.37	51.2	80.55
Al_2O_3	21.50	18.90	21.18	15.60	17.44	18.03	15.1	14.1	21.77	0.76	21.9	10.23
Fe_2O_3	2.72	1.24	3.23	1.09	2.19	1.28	6.07	0.70	2.64	0.36	2.40	1.25
FeO	6.04	4.24	4.16	3.98	6.90	8.15	7.50	11.5	4.10	0.70	2.76	0.08
CaO	0.98	3.30	0.79	2.42	11.05	9.82	8.73	10.5	1.39	51.0	0.40	0.06
MgO	3.00	1.64	2.75	6.80	7.79	8.45	4.50	6.22	2.74	0.59	5.92	0.48
Na_2O	1.43	2.17	0.90	1.70	1.96	2.08	2.09	1.97	1.32	0.04	5.14	2.03
K_2O	3.01	1.88	2.30	4.92	0.48	0.60	0.88	0.61	2.20	0.30	2.63	4.38
H_2O^+	3.53	3.56	2.61	1.11	0.53	1.52	1.73	1.04	1.23	0.41	3.93	0.94
H_2O^-	0.02	0.05		0.05	0.22	0.23	0.15	0.09	0.07		0.23	
P_2O_5	0.11	0.11	0.09	0.12	0.07	0.07	0.09	0.08	0.12	0.22	0.14	0.002
TiO_2	0.98	0.64	0.74	0.53	0.30	0.28	0.98	1.10	0.88	0.05	0.95	0.09
Cr_2O_3	0.02		0.01		0.09	0.03						
MnO	0.05	0.05	0.04	0.03	0.12	0.14	0.17	0.24	0.06	0.32	0.07	0.00
V_2O_3	0.01		0.01		0.07	0.04						
Co_2	0.12		0.20	0.82	0.39	0.33	0.25	0.14	0.53	40.0	0.06	
S				0.27	0.08	0.09	0.08	0.01		0.06	0.01	
Total	98.59	99.78	99.74	98.70	99.57	99.39	100.07	99.5	100.6	99.4		

1. Low greenschist facies argillaceous metasediment (chlorite zone) McKim Formation. Muscovite-chlorite-quartz-albite-sphene-ilmenite (Card, 1964).
2. Upper greenschist facies argillaceous metasediment (garnet zone) McKim Formation. Muscovite-quartz-oligoclase-chlorite-chloritoid-garnet-ilmenite (Card, 1976c).
3. Amphibolite facies argillaceous metasediment (staurolite zone) McKim Formation. Staurolite-biotite-garnet-andesine-quartz-muscovite-rutile-ilmenite (Card, 1964).
4. Scapolitic metasediment (biotite zone) Espanola Formation. Quartz-scapolite-muscovite-biotite-chlorite-calcite-iron oxides-pyrite (Card, 1976c).
5. Unmetamorphosed pyroxene gabbro (Nipissing). Labradorite-orthopyroxene-clinopyroxene-ilmenite-magnetite (amphibole biotite, chlorite, sulphide) (Card and Pattison, 1973).
6. Low greenschist facies metagabbro (Nipissing). Albite-actinolite-quartz-biotite-chlorite-epidote-sphene-ilmenite-magnetite (Card and Pattison, 1973).
7. Amphibolite facies metagabbro (Nipissing). Andesine-hornblende-quartz-biotite-sphene-ilmenite-magnetite (Card and Pattison, 1973).
8. Amphibolite facies metabasalt, Elsie Mountain Formation. Hornblende-plagioclase-quartz-biotite-chlorite-epidote-iron-titanium oxides (Card, 1976c).
9. Argillaceous metasediment, amphibolite facies, Stobie Formation. Staurolite-biotite-chlorite-quartz-plagioclase-sphene-ilmenite (Card, 1976c).
10. Marble, lower greenschist facies, Espanola Formation. Calcite- (quartz-muscovite) (Card, 1976c).
11. Argillaceous metasediment, upper greenschist facies, Gowganda Formation. Biotite-garnet-muscovite-quartz-plagioclase-iron-titanium oxides-sphene (Card, 1976c).
12. Aluminous metasandstone, upper greenschist facies, Lorrain Formation (Young, 1973).

Table 3

Spectrographic and microprobe analysis of metamorphic minerals from mafic meta-igneous and argillaceous metasedimentary rock of the Sudbury-Espanola area

	Hornblende	Actinolite	Chlorite		Biotite		Garnet		
	C-35	CD-130	C-58A	W-1	SA-1	C-64	SA-1	C-64	C-35
SiO_2				22.42				35.0	
Al_2O_3				26.47				21.3	
Fe_2O_3	1.2				1.78	3.45			
FeO	16.2	9.82	26.8	26.35	16.8	13.8	33.5	34.8	31.1
Total Fe	13.5	7.45	20.7		14.4	13.1	26.0	27.1	24.2
MgO	5.03	14.6	10.6	13.83	10.6	11.0	2.31	2.59	2.31
CaO	9.02	12.9	0.04	0.03	0.37	0.21	2.92	3.15	4.87
Na_2O	1.09	0.31	0.13		0.46	0.24			
K_2O	0.24	0.11	0.11	0.02	5.64	7.17			
TiO_2	2.27	0.30	0.37	0.07	1.79	1.53	0.13	0.14	0.18
MnO	0.19	0.20	0.15	0.13	0.05	0.03	1.03	0.92	0.95
ZnO									

	Staurolite			Chloritoid	Andalusite	Ilmenite	Rutile
	SA-1	C-64	W-1	W-1	W-1	W-1	W-1
SiO_2			27.68	24.84	37.59	0.05	0.13
Al_2O_3			55.82	40.65	62.58	0.11	0.05
Fe_2O_3							
FeO	13.6	13.2	12.26	23.46	0.28	46.21	0.25
Total Fe	10.6	10.3					
MgO	1.72	2.03	1.17	3.09	0.00	0.00	
CaO	0.10	0.07	0.00	0.00	0.00	0.07	
Na_2O	0.09	0.07			0.00		
K_2O	0.16	0.08	0.00				
TiO_2	0.53	0.51	0.70	0.00	0.02	51.63	98.80
MnO	0.05	0.05	0.26	0.45		2.18	
ZnO			0.70	0.00			

Host Rock Characteristics:

C-35-garnet amphibolite (meta-Nipissing Diabase): andesine-13%; quartz-4%; hornblende-64%; garnet-3%; epidote, chlorite-3%; magnetite-ilmenite, sulphides-13% (Card, 1964).

CD-130-actinolite amphibolite (meta-Nipissing Diabase): actinolite-98.6%; quartz-0.4%; ilmenite-1.0% (Card, 1964).

C-58A-garnet-chlorite argillaceous metasediment; plagioclase-20%; quartz-50%; muscovite-22%; chlorite-7%; garnet-0.4%; others-0.6% (Card, 1964).

W-I-argillaceous metasediment; staurolite-chloritoid-andalusite-plagioclase (An_{27})-muscovite (Mu_{72})-quartz-rutile-ilmenite-pyrrhotite (Fox, 1971).

SA-1-biotite-garnet-staurolite argillaceous metasediment: plagioclase (oligoclase)-14%; muscovite-17%; biotite-8%; quartz-30%; garnet-12%; staurolite-18%; ilmenite, sphene-1% (Card, 1964).

C-64-biotite-garnet-staurolite argillaceous metasediment: plagioclase (An_{26})-15%; quartz-29%; muscovite-15%; biotite-10%; garnet-1%; staurolite-28%; others-1% (Card, 1964).

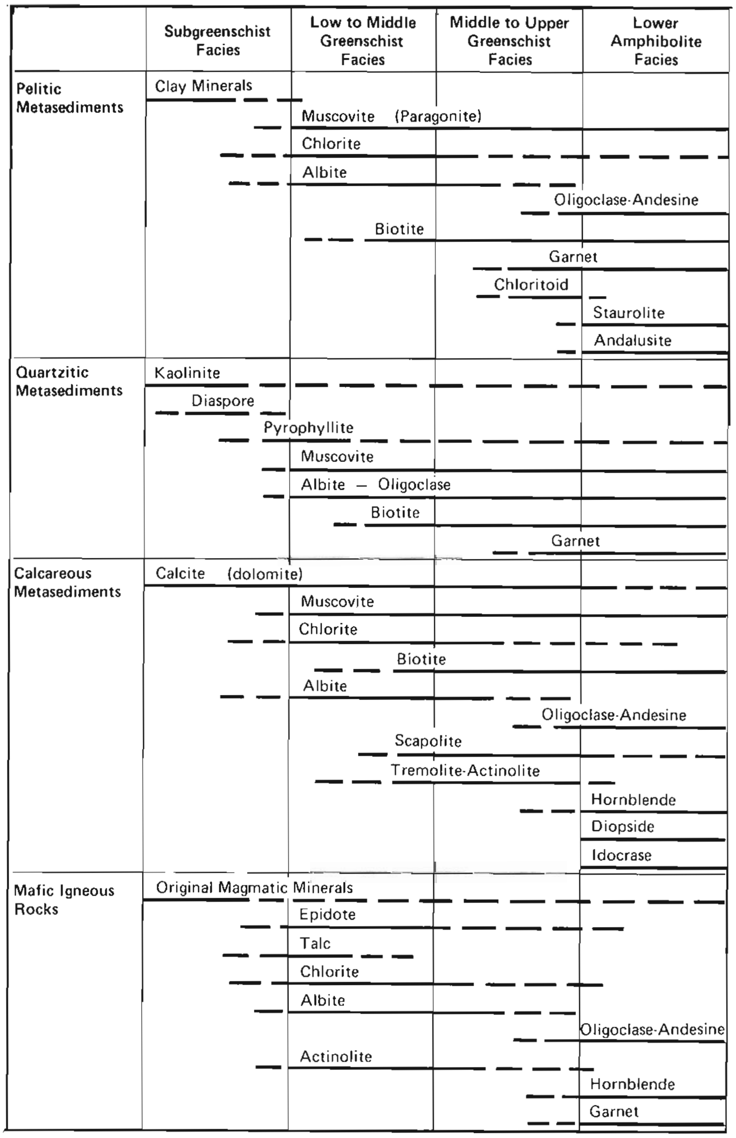

Figure 2. Metamorphic minerals present in the major rock types of the eastern Southern Province at various metamorphic grades.

Argillaceous rocks contain:

chlorite — muscovite - albite - quartz

muscovite — albite - quartz

biotite — quartz

Mafic igneous rocks are commonly only partly altered and, in addition to original pyroxene, amphibole, and plagioclase, contain the following:

actinolite — albite

chlorite — biotite - quartz - albite

chlorite — amphibole - albite

hornblende — biotite - albite

Quartz, epidote, carbonate, muscovite, sphene, iron oxides, and sulphides are prevalent and many rocks contain both actinolitic amphibole and blue-green hornblende.

Scapolite in the calcareous metasediments of this and higher grade metamorphic zones form inclusion-filled porphyroblasts that tend to be euhedral in the marble (Fig. 4) and anhedral in the metamorphosed calcareous siltstone and wacke (Fig. 5). Although the scapolite occurs in Ca-rich rocks, it is sodic (marialite-rich); some scapolite porphyroblasts have sodic cores and calcic rims (M.J. Frarey, pers. comm., 1977). Chemical analyses show that scapolitic rocks contain nearly twice as much Na_2O as otherwise comparable non-scapolitic rocks (Card, 1976a). It is probable that the soda required to form scapolite was originally present in the rocks in evaporite minerals such as halite.

Middle to Upper Greenschist Facies

Rocks of the middle to upper greenschist facies (garnet and chloritoid zones) are recrystallized with generally well preserved primary textures and structures. Quartzitic rocks display the following assemblages:

quartz — biotite - muscovite

quartz — biotite - muscovite - plagioclase

quartz — biotite - muscovite - plagioclase - epidote

quartz — muscovite - biotite - garnet - plagioclase

In addition to the foregoing there are aluminosilicate-bearing assemblages developed in quartz-rich metasandstone of the Lorrain Formation along the north shore of Lake Huron. These include:

quartz — kyanite - muscovite (paragonite) ± pyrophyllite ± kaolinite

quartz — andalusite - kyanite - muscovite (paragonite) ± pyrophyllite ± kaolinite

Minor amounts of dumortierite, a pink aluminum borosilicate, tourmaline, and fuchsite, a green chrome mica, are also present in these rocks.

Kyanite (Fig. 6), andalusite (Fig. 7), and pyrophyllite (Fig. 8) have developed from kaolinite which in turn was probably formed by weathering and authigenic alteration of feldspars. On the basis of the assemblages in the sandstone and in nearby argillaceous metasediments, the grade of metamorphism appears to be upper greenschist facies. The textural evidence, notably intergrowths of kyanite and andalusite (Fig. 7), indicates that the two minerals were formed at the same time, not at different times under different stress conditions as postulated by Church (1967). Possibly these minerals owe their formation to a combination of primary materials (kaolinite and quartz), sluggish reactions near the triple point in the aluminosilicate system, and high confining pressure existent in the axial zone of a major isoclinal fold, the La Cloche Syncline (Fig. 1). Calcareous rocks contain:

scapolite — calcite (dolomite)

scapolite — quartz - muscovite - biotite - chlorite - calcite (dolomite)

scapolite — quartz - biotite - muscovite - plagioclase tremolite - actinolite - biotite ± scapolite - calcite (dolomite)

epidote — quartz - plagioclase

actinolite — epidote - plagioclase - microcline - quartz

hornblende — biotite ± scapolite - quartz

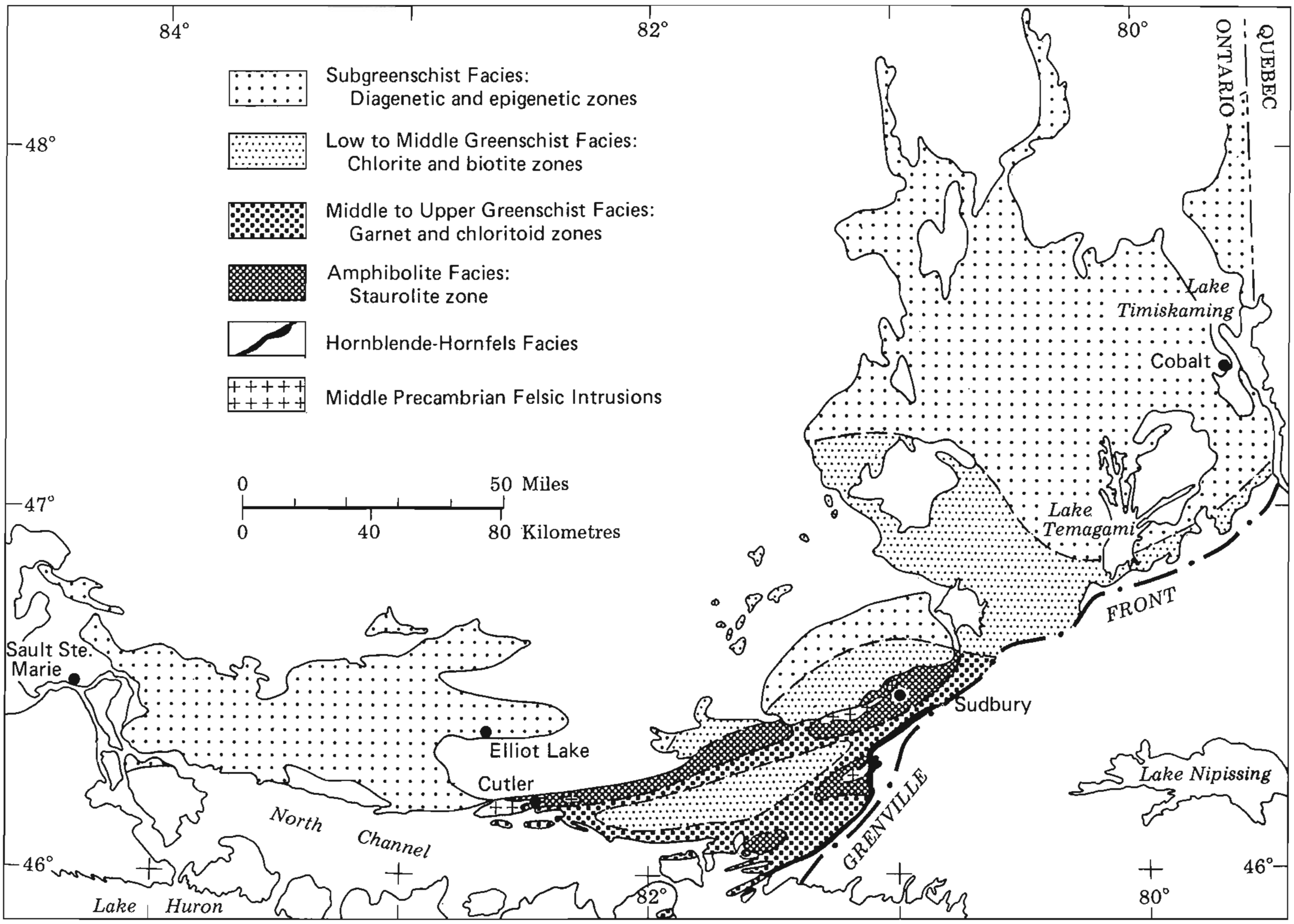

Figure 3. Patterns of metamorphism in the eastern part of the Southern Province.

Argillaceous mineral assemblages include:

biotite — muscovite - plagioclase - quartz

biotite — garnet - muscovite

muscovite — chloritoid - chlorite - plagioclase - quartz

muscovite — chloritoid - garnet - plagioclase - quartz

Mafic meta-igneous rocks contain:

hornblende — plagioclase

hornblende — biotite - plagioclase

hornblende — biotite - garnet - plagioclase

Quartz, epidote, chlorite, sphene, ilmenite, magnetite, and sulphides are commonly present in all of the foregoing rocks. The plagioclase is generally oligoclase $An_{25} \pm _5$.

Amphibolite Facies

Rocks of amphibolite facies, although recrystallized, commonly retain primary structures and textures. Quartzitic rocks display the following assemblages:

biotite - muscovite - quartz

biotite - muscovite - plagioclase - quartz

biotite - muscovite - garnet - calcite (dolomite) - quartz

Calcareous rocks display:

biotite - muscovite ± scapolite - plagioclase ± quartz ± epidote ± calcite (dolomite)

hornblende - biotite ± scapolite - plagioclase ± quartz ± epidote ± calcite (dolomite)

diopside - plagioclase - microcline - quartz

pyroxene (augite?) - microcline - actinolite - epidote - quartz ± calcite (dolomite)

idocrase - diopside - garnet - calcite (dolomite) - quartz

Argillaceous assemblages include:

staurolite - chloritoid - chlorite - muscovite (paragonite) - plagioclase - quartz

staurolite - chloritoid - chlorite - andalusite - muscovite (paragonite) - plagioclase - quartz

staurolite - chlorite - andalusite - muscovite - plagioclase - quartz

staurolite - garnet - biotite - muscovite - plagioclase - quartz

andalusite - biotite - muscovite - plagioclase - quartz

garnet - muscovite - biotite - plagioclase - quartz

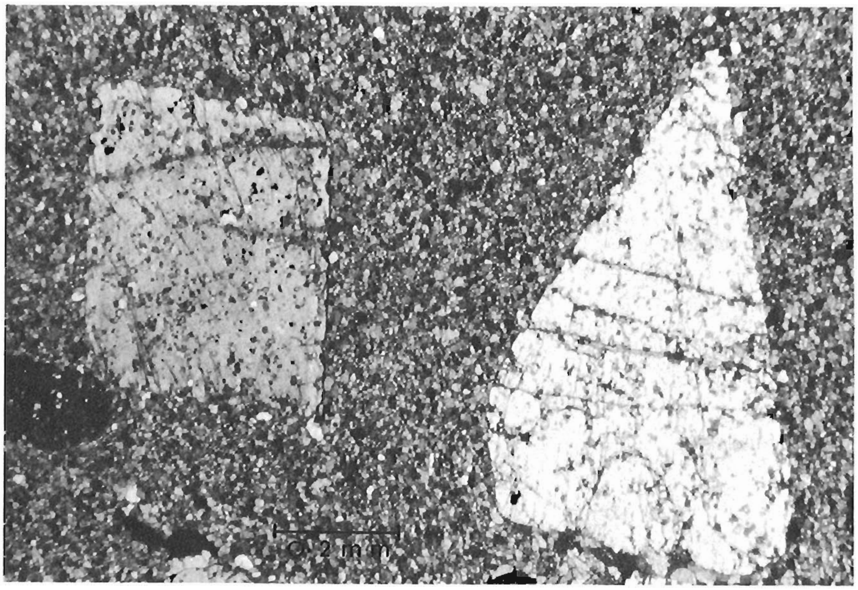

Figure 4

Euhedral scapolite porphyroblasts in marble, Espanola Formation, middle greenschist facies, Sudbury-Manitoulin Area. (GSC 203218-E)

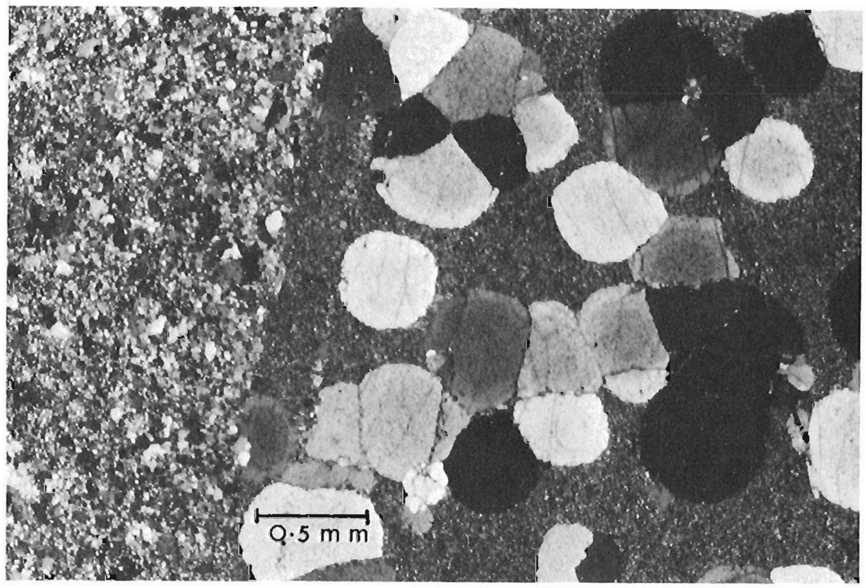

Figure 5

Anhedral scapolite porphyroblasts in calcareous siltstone-sandstone of the Espanola Formation, middle greenschist facies, Sudbury-Manitoulin area. (GSC 203218-D)

Mafic meta-igneous rocks contain:

hornblende - plagioclase

hornblende - biotite - plagioclase

hornblende - biotite - garnet - plagioclase

In addition sphene, rutile, chlorite, ilmenite, magnetite, and sulphides are commonly present. The plagioclase in the amphibolite facies rocks is mainly andesine. Chemical analyses of metamorphic minerals from mafic meta-igneous and argillaceous metasedimentary rocks are given in Table 3.

The foregoing metamorphic mineral assemblages, notably the coexistence of staurolite and andalusite (Fig. 9), indicate that the regional metamorphism occurred under pressure-temperature conditions corresponding to the low pressure intermediate type of Miyashiro (1961). The local occurrence of kyanite (Fig. 6) may indicate locally higher pressures, possibly tectonically induced. Fox (1971) estimated that the chloritoid-staurolite transition, marking the change from the greenschist to the amphibolite facies in an area near Sudbury, occurred at temperatures of 500°C to 550°C and pressures of 4.45 to 4.75 kb[1].

In Figure 2 the characteristic metamorphic minerals that exist in each of the major rock types in the various metamorphic zones are shown. The transition from the subgreenschist to the greenschist facies is marked mainly by the appearance of pyrophyllite, muscovite, and chlorite at the expense of clay minerals such as illite-montmorillonite and kaolinite. The transition from the middle to the upper greenschist facies is not well defined, but is generally marked by the appearance of oligoclase at the expense of albite and the appearance of garnet, chloritoid and actinolitic amphiboles in rocks of appropriate composition. The transition from the greenschist to the amphibolite facies is recorded in aluminous, iron-rich argillaceous metasediments by the incoming of staurolite at the expense of chloritoid (Fig. 10). In addition, argillaceous rocks of the amphibolite facies commonly display well developed porphyroblasts of

[1] 1 kb = 10^5 kPa.

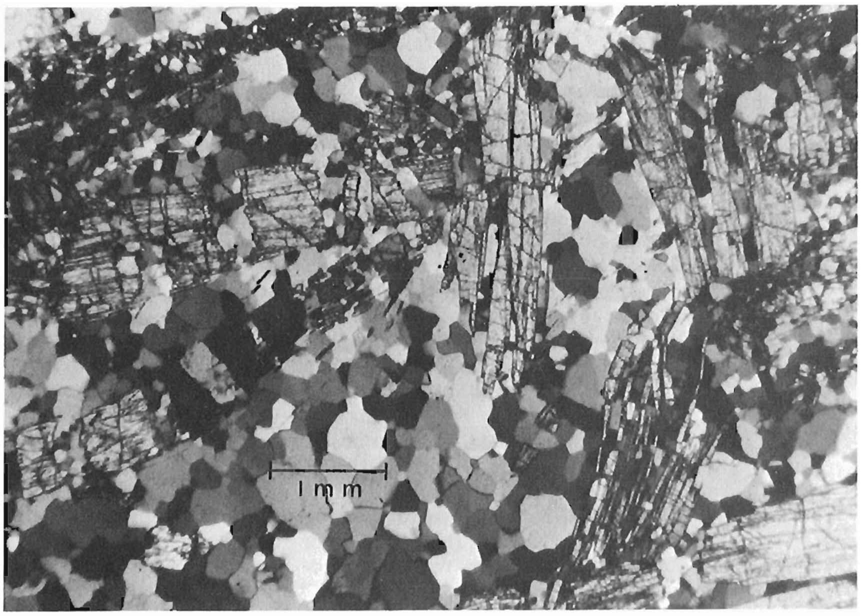

Figure 6

Kyanite porphyroblasts in aluminous orthoquartzite of the Lorrain Formation, upper greenschist facies, Sudbury-Manitoulin area. Note disruption of the kyanite crystals. (GSC 203218-T)

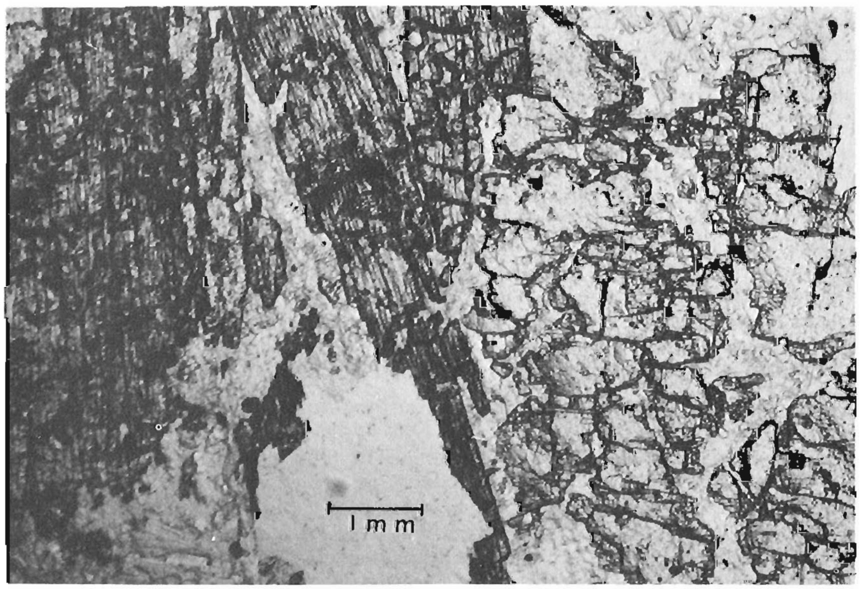

Figure 7

Kyanite – andalusite intergrowths in aluminous orthoquartzite of the Lorrain Formation, upper greenschist facies, Sudbury-Manitoulin area. (GSC 203218)

andalusite, garnet (Fig. 11), and intermediate plagioclase (oligoclase, andesine). The calcareous metasediments contain hornblende, and locally, diopside and idocrase, whereas the mafic igneous rocks consist essentially of hornblende and intermediate plagioclase with garnet developed locally.

The contact metamorphic aureole adjacent to the Grenville Front felsic plutons is approximately 600 m wide and is superimposed on the previously regionally metamorphosed rocks of the Huronian Supergroup. The effects of the contact metamorphism include development of spotted hornfels textures with growth of metamorphic porphyroblasts of chlorite, biotite, alkali feldspars, andalusite, and cordierite. Typical assemblages in argillaceous metasediments within the aureole include:

biotite - chlorite - muscovite - plagioclase - quartz

andalusite - biotite - muscovite - plagioclase ± microcline - quartz

cordierite - andalusite - biotite - microcline - plagioclase - quartz

These assemblages are generally diagnostic of the hornblende hornfels facies of contact metamorphism.

PATTERNS OF REGIONAL METAMORPHISM AND RELATIONSHIP TO MIDDLE PRECAMBRIAN DEPOSITION, DEFORMATION, AND PLUTONIC ACTIVITY

The eastern part of the Southern Province can be divided into several areas displaying differences in intensity of regional metamorphism. The variations in metamorphic grade are matched in a general way, by variations in thickness and facies of the Middle Precambrian supracrustal rocks, by variations in intensity and style of deformation, and by variations in Middle Precambrian felsic plutonic activity. The subdivisions, shown in Figures 1 and 3, include the Sault Ste. Marie-Elliot Lake area, the Cobalt Embayment, the Sudbury Structure, and the Sudbury-Cutler area. For the most part, these areas are separated by fault/fold lines that mark relatively sharp discontinuities in structural style and intensity of deformation and metamorphism. These same discontinuities also mark relatively abrupt changes in thickness of the Huronian sequence, and to some extent, variations in facies.

In the Sault Ste. Marie-Elliot Lake area, the Huronian sequence is relatively thin (ca. 4500 m) and consists predominantly of quartz and feldspar-rich clastic sediments. The strata have been mildly deformed, tilted, and folded into open concentric folds with gentle, variable plunges. Minor tectonic structures are only locally developed. Middle Precambrian felsic plutons are absent. The rocks of this area have been little metamorphosed under conditions herein termed "subgreenschist facies". They contain original detrital minerals, such as plagioclase, chemically precipitated minerals, such as anhydrite, minerals of probable diagenetic origin such as kaolinite and illite-montmorillonite, and low grade metamorphic minerals such as stilpnomelane, chlorite, muscovite, diaspore, and pyrophyllite. Andalusite has been reported locally (Chandler, 1969; Wood, 1973). Tectonic foliations are defined by oriented muscovite and chlorite, and in some areas the Nipissing Diabase and Early Precambrian granitic rocks have been retrogressively altered.

The Huronian sequence of the Cobalt Embayment is also relatively thin (4500 to 6000 m) and consists predominantly of clastic metasediments. Huronian strata of the eastern part of the Cobalt Embayment are almost undeformed whereas those of the western part are moderately folded, forming broad, open, doubly-plunging domes and basins. The structure of the western margin of the Cobalt Embayment is complex and there is evidence for sliding of large blocks of supracrustal rocks under the influence of gravity (Card et al., 1972). The patterns of regional thermal metamorphism generally conform to the structural variations. In the east, the Huronian rocks are apparently unmetamorphosed for the most part, but locally contain porphyroblasts of chlorite and muscovite. They have probably been altered under subgreenschist facies conditions. In the central part of the Cobalt Embayment, biotite, chlorite, and pyrophyllite of metamorphic origin are present, and in the western part mineral assemblages indicative of low greenschist facies (chlorite and biotite zones) are prevalent (Card et al., 1970; Meyn, 1973). Middle Precambrian felsic plutons are absent.

The rocks of the Sudbury Structure, including the Whitewater Group and the Sudbury Nickel Irruptive, have been deformed and metamorphosed. The rocks of the northern part of the Sudbury Structure however are essentially unmetamorphosed. The primary mineralogy of the Nickel Irruptive has been largely preserved and delicate glass shards in the Onaping Formation, although devitrified, have survived. In contrast, the rocks of the southern part of the Structure have been relatively highly deformed and mildly metamorphosed. Mineral assemblages indicative of low greenschist facies conditions are present in the Whitewater Group in the south and alteration of the primary minerals of the Nickel Irruptive is common (Card, 1968; Rousell, 1975). In the rocks of the Whitewater Group, deformation and regional metamorphism were essentially contemporaneous as oriented metamorphic minerals define the prominent foliation and lineation. The metamorphism that affected the rocks of the Sudbury Structure, however, is of a distinctly lower grade than that which affected the Huronian rocks immediately to the south. This, and the radiometric age data which suggest that the main culmination of regional metamorphism occurred prior to emplacement of the Nickel Irruptive, indicate that the metamorphism that affected the Sudbury Structure is somewhat younger than the major metamorphism of the Huronian rocks. This will be further discussed under "Ages of Metamorphism".

The rocks of the Sudbury-Cutler area (Fig. 1) are relatively highly metamorphosed and show the effects of regional metamorphism under conditions ranging from low greenschist to low amphibolite facies. It is also in this area that the Huronian sequence reaches its greatest thickness (ca. 11 000 m) and contains significant amounts of volcanic rocks and greywacke turbidites. The rocks are relatively highly deformed with generally tight folds and relatively well developed foliations and lineations. Middle Precambrian (Aphebian) felsic plutons, though not abundant, are present and are situated in areas of higher grade metamorphism. Studies of the distribution of metamorphic mineral assemblages in the Sudbury-Cutler area show that there are two northeast-trending belts of higher grade rocks (middle greenschist to lower amphibolite facies) superimposed on a lower grade (low to middle greenschist facies) terrane. The highest grade rocks (low amphibolite facies) are distributed within these belts in the form of elliptical "nodes" (Fig. 3). The higher grade assemblages occur south of the Murray Fault and branch faults that traverse the Sudbury Structure. The northern high grade metamorphic belt coincides with a major structural element, the Baldwin Anticlinorium, while the southern belt is approximately parallel with the Grenville Front and corresponds to a series of major anticlinoria in the Huronian rocks, the McGregor Bay and Lake Panache structures. The metamorphic "lows" to the north and between the two high grade belts correspond to major synclinoria, the Porter and Bass Lake structures.

There is a general spatial relationship between areas of higher grade metamorphism and Middle Precambrian felsic intrusions. However, there are inconsistent temporal relationships between the times of emplacement of these plutons and the regional metamorphism. The Murray and Creighton bodies have been deformed and metamorphosed along with the surrounding Huronian country rocks whereas the Cutler and Grenville Front plutons cut previously metamorphosed and deformed Huronian rocks. Major metamorphism occurred after emplacement of the Murray and Creighton plutons (2100-2200 Ma) and before intrusion of the Cutler (1750 Ma) and Grenville Front (1600-1730 Ma) plutons.

AGES OF METAMORPHISM

Studies of the relationships between structural elements such as foliations and lineations and metamorphic porphyroblasts in the Middle Precambrian (Aphebian) rocks by Card (1964) indicate the following sequence of events:

1) Early deformation and growth of "low grade" minerals such as muscovite and chlorite.

2) A metamorphic culmination, probably about 1900 Ma ago, with equilibration of the metamorphic assemblages, and, under appropriate conditions, growth of "high grade" minerals such as biotite, garnet, staurolite, andalusite, and kyanite. The metamorphic culmination was approximately accompanied by secondary deformation involving major folding in the areas of highest grade metamorphism. There is evidence for prekinematic, synkinematic, and postkinematic growth of porphyroblasts with respect to the secondary deformation.

3) Late, postmetamorphic deformation involving formation of strain-slip cleavages, cross-folds, and destruction of metamorphic porphyroblasts.

Metamorphosed rocks of the eastern Southern Province commonly yield Rb-Sr and K-Ar radiometric dates in the range 1600 to 1900 Ma, the approximate age of the Hudsonian Orogeny (Stockwell et al., 1970) and the Penokean Orogeny (Goldich, 1968). Fairbairn et al. (1969) have estimated a mean Rb-Sr isochron age of 1950 ± 100 Ma for major metamorphism of the Huronian of the Sudbury-Espanola area. Van Schmus (1976) has recently placed the age of the Penokean Orogeny in the west at approximately 1900 Ma.

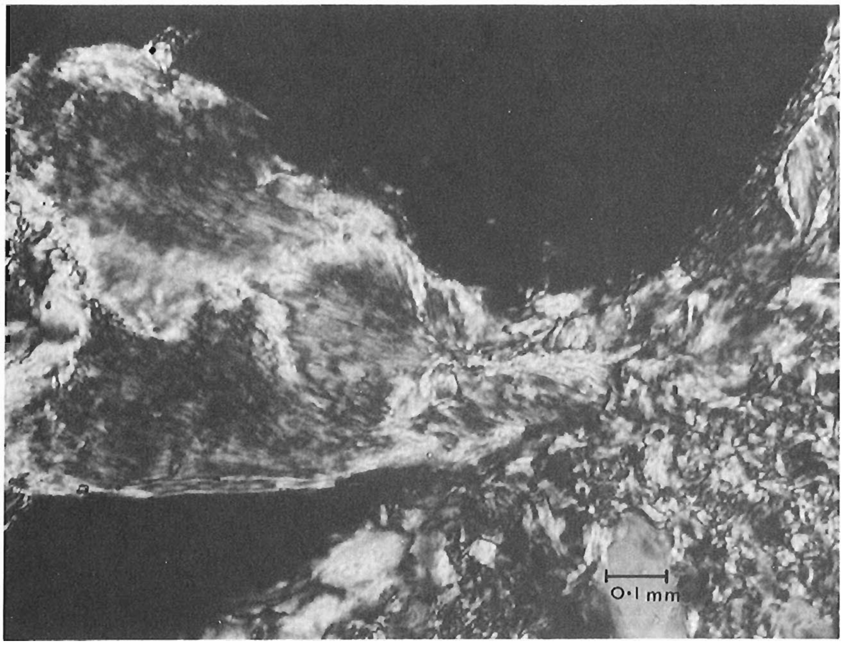

Figure 8

Pyrophyllite in aluminous orthoquartzite of the Lorrain Formation, subgreenschist facies, Maple Mountain area. (GSC 203218-M)

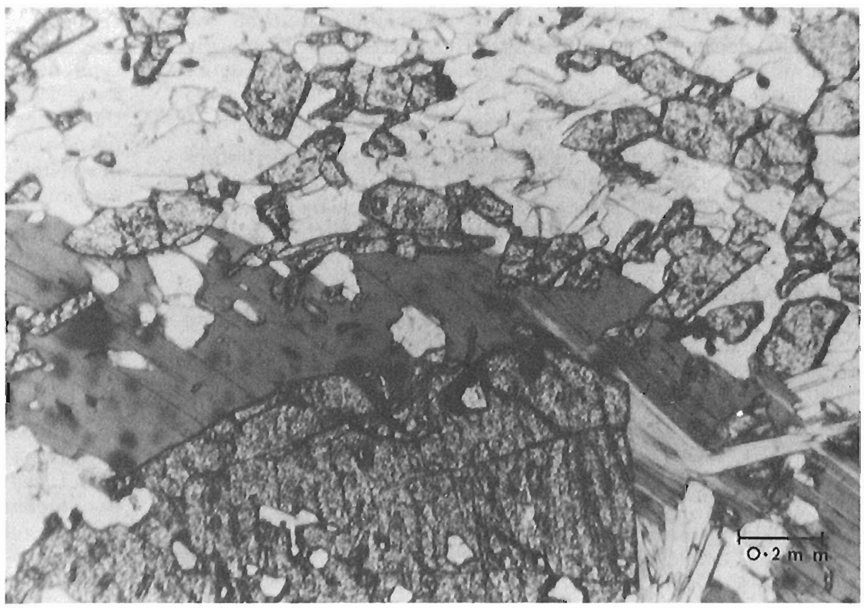

Figure 9

Andalusite (bottom), biotite (middle), and staurolite (top) in pelitic metasediment of the McKim Formation, amphibolite facies, Sudbury-Manitoulin area. (GSC 203218-U)

Consequently, it appears that the entire Southern Province was affected by regional metamorphism and deformation approximately 1900 Ma ago (during the Penokean Orogeny).

There is evidence for deformation, metamorphism, and felsic plutonic activity throughout the Southern Province preceding emplacement of the Nipissing Diabase some 2150 Ma ago. Early deformation and low grade metamorphism of the Huronian rocks and emplacement of the Creighton and Murray plutons, dated at about 2200 Ma (Gibbins et al., 1972), probably occurred during this event. Field evidence for pre-Nipissing deformation of the Huronian rocks has been presented by Card (1968), and others and consists essentially of the following:

- Some Nipissing Diabase intrusions transect the axes of folds such as the McGregor Bay Anticline and Vernon Syncline in the Sudbury-Cutler area and numerous folds in the Cobalt Embayment.
- Nipissing Diabase intrusions transect early formed foliation related to major fold structures that are present in the Huronian country rocks but not in the intrusions themselves.

Van Schmus (1976) presented geochronological evidence for this early orogenic event and stated: "It is represented by at least some deformation of Huronian strata and emplacement of Nipissing Diabase in Ontario, by metamorphism, granitic plutonism, and deformation in Michigan, and by emplacement of mafic dykes in Minnesota. This orogenic pulse is distinct from and older than that involving Early Proterozoic strata in Michigan, Minnesota and Wisconsin for which the term Penokean Orogeny is more properly applied".

There is also paleomagnetic evidence indicating pre-Nipissing Diabase deformation. Morris (1977) states: "Together with other presently available Aphebian paleomagnetic results these data indicate that an event of folding and subsequent remagnetization affected the whole region from Quebec to Minnesota prior to intrusion of the Nipissing Diabase at approximately 2150 m.y. B.P. ..."

After emplacement of the Sudbury Nickel Irruptive and the Whitewater Group there was deformation and low grade metamorphism of the rocks of the Sudbury Structure. Huronian rocks display the effects of this metamorphism in the form of extensive retrograde alteration of the pre-existing high grade metamorphic mineral assemblages. Van Schmus (1976) has presented evidence for an extensive thermal-deformational event that affected large areas in the western Southern Province some 1600 to 1700 Ma ago. It is possible that the post-Nickel Irruptive deformation and metamorphism in the east are correlative with this event and may be correlative with the Hudsonian Orogeny of the Churchill Province.

Brocoum and Dalziel (1974) have proposed that deformation of the Sudbury Nickel Irruptive and Whitewater Group, the eastern part of the Southern Province, and the adjacent Grenville Front Tectonic Zone occurred during a single event, the Penokean Orogeny. By inference, the regional metamorphism would also be of a single, Penokean age. Brocoum and Dalziel's hypothesis is untenable for a number of reasons including the following:

- There is geological, geochronological and paleomagnetic evidence for a widespread deformational-metamorphic-plutonic event prior to emplacement of the Nipissing Diabase some 2150 Ma ago (Card, 1968; Church, 1968; Van Schmus, 1976; Morris, 1977).
- The available geochronological data indicate that the main metamorphic culmination occurred approximately 1900 Ma ago, before the emplacement of the Sudbury Nickel Irruptive some 1840 Ma ago (Krogh and Davis, 1974).
- The Huronian rocks adjacent to the Sudbury Structure bear evidence of earlier deformation and metamorphism, in the form of structural elements and metamorphic mineral assemblages that the rocks of the Sudbury Structure do not have.
- Felsic plutons along and within the Grenville Front Tectonic Zone were emplaced 1600 Ma and 1730 Ma ago (Krogh et al., 1971). These plutons cut pre-existing structures in the Huronian rocks and have superimposed contact metamorphic aureoles on the previously regionally metamorphosed Huronian rocks. The plutons are themselves deformed and metamorphosed within the Grenville Front Tectonic Zone as are even younger intrusions (Card et al., 1975; Card, 1976b; Lumbers, 1975).

Clearly, the last major deformation and metamorphism in the Grenville Front Tectonic Zone was distinctly younger than that in the adjacent Southern Province. It is equally clear that the rocks of the eastern Southern Province were subjected to more than one deformational-metamorphic event, some predating, others postdating the Sudbury Structure.

Wynne-Edwards (1972) has described the eastern part of the Southern Province as part of a "foreland" zone of the Grenville orogenic belt. He attributes the northeast-trending deformation of the Huronian sequence and the Sudbury Structure to the Grenville Orogeny, and interprets the metamorphic patterns in the Southern Province as resulting from passive uplift and deformation of subhorizontal "fossil isograds" during the Grenville Orogeny. The rocks of the eastern Southern Province have been affected by Grenville deformation, but only in a restricted zone no more than a few kilometres in width adjacent to the Grenville Front (Card, 1976b,c; Lumbers, 1975). The deformational events that affected the remainder of the Southern Province occurred prior to emplacement of the Grenville Front granites and the Cutler pluton some 1700 Ma ago and consequently, had nothing to do with the 1000 to 1200 Ma Grenville Orogeny. Similarly, metamorphic grades in the Southern Province rocks do not increase uniformly toward the Grenville Front, nor do the patterns of metamorphic isograds conform to the structural-stratigraphic trends in the Huronian rocks as would be demanded by Wynne-Edwards model of deformed "fossil isograds". The metamorphic patterns in the eastern Southern Province are similar to those in the west, in areas far removed from the Grenville influence. Cannon (1973) concluded that since there has been no significant postmetamorphic tilting or other deformation in the west, the metamorphic isograds are approximately isotherms and the present erosion surface is an isobaric surface. The same is probably true in the east.

Conversely, it is probable that the northwestern Grenville Province was affected by Middle Precambrian deformation and metamorphism. There is field evidence for early, pre-Grenville deformation and metamorphism within the Grenville Front Tectonic Zone (Lumbers, 1975) and Krogh and Davis (1969) have presented isotopic evidence for major metamorphism approximately 1800 Ma ago in the northwest Grenville. In an area south of Sudbury and northwest of the Grenville Front, the Huronian rocks are migmatitic and may represent a sample of the Middle Precambrian high-rank metamorphism that was concentrated mainly in the adjacent Grenville Province.

SUMMARY AND CONCLUSIONS

Major metamorphism of the low-pressure-intermediate type affected the rocks of the eastern Southern Province some 1900 Ma ago during the Penokean Orogeny. In addition, there was earlier, pre-Nipissing Diabase deformation and low grade metamorphism approximately 2200 Ma ago, and later, post-Nickel Irruptive deformation and low grade metamorphism some 1600-1700 Ma ago.

Major metamorphism was approximately accompanied by secondary deformation, and there is general coincidence of areas of highest grade metamorphism with the thickest accumulations of Huronian cover rocks, with structural elements and trends, with felsic plutons, and with zones of greatest secondary deformation. However, in detail, isograds transect fold axes, indicating that the higher grade metamorphism was superimposed on previously folded rocks or outlasted the major phases of folding. There are great variations in the degree of development of tectonic structures and metamorphic porphyroblasts, even in the zones of highest grade metamorphism. There is little correspondence between style or intensity of deformation and metamorphic grade. Rocks with weakly developed tectonic foliations commonly display metamorphic mineral assemblages corresponding to the amphibolite facies. Well foliated rocks may contain assemblages varying from low greenschist to low amphibolite facies. Deformation and thermal metamorphism were relatively independent of one another in detail.

The distribution of metamorphic mineral assemblages indicates that the distribution of temperature was notably irregular. There was a broad thermal high within the belt as a whole with elongate, higher temperature zones along the structural axes of the belt, and high temperature "nodes" within the surrounding lower grade terrane. These metamorphic variations are not attributable to variations in rock composition or fabric and hence must be due to variations in heat supply. The direction of elongation of higher temperature zones is generally parallel with the foliation, and thus anisotropic heat conduction could account for the patterns. The lowest grade metamorphic zones are more extensive than those of intermediate and higher grade. Calculations based on the assumption that the metamorphic isograds approximate isotherms indicate geothermal gradients

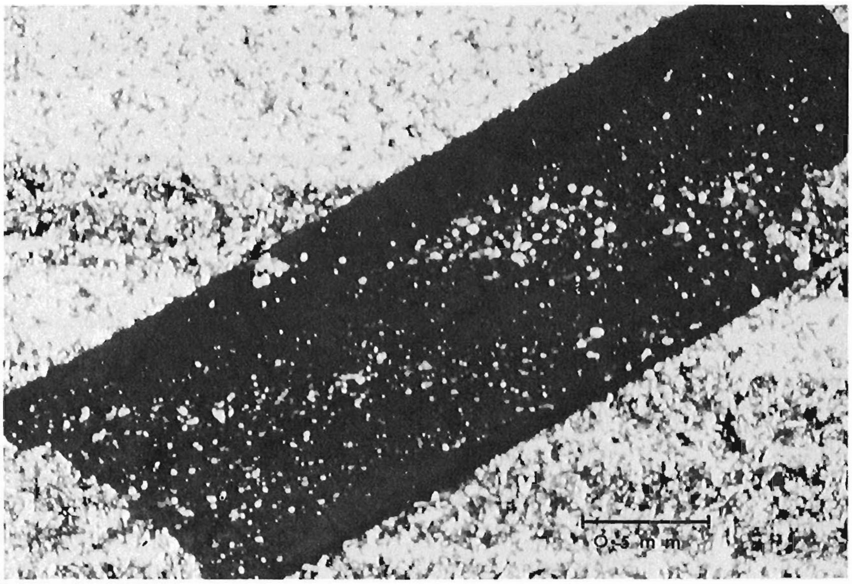

Figure 10. Chloritoid porphyroblast in pelitic metasediment of the McKim Formation, upper greenschist facies, Sudbury-Manitoulin area. Note non-rotated inclusion trains in the chloritoid. (GSC 20321-F)

Figure 11. Garnet, biotite, and muscovite in pelitic metasediment of the McKim Formation, Sudbury-Manitoulin area. Note rotated inclusion trains in the garnet. (GSC 20321-H)

on the order of 70°C to 150°C per km (James, 1955; Card, 1964). The higher grade nodes must represent heat "pipes" through which thermal conduction was relatively rapid. This implies that at the crustal levels (12 to 17 km) represented by the Southern Province, regional isograds and paleoisotherms are relatively steeply dipping surfaces.

The general correspondence between variations in metamorphic grade, intensity of deformation, felsic plutonic activity, and original variations in thickness and facies of the cover rocks indicates that all are genetically related. In the case of the Southern Province, these variations are probably related to differences in the behaviour of the basement upon which the supracrustal sequences were deposited. In the north, the Early Precambrian basement remained relatively stable, both during deposition of the Huronian rocks and later, during Middle Precambrian orogenic events. In contrast, the basement beneath the southern part of the belt, presumably also Early Precambrian crustal material, was relatively unstable during deposition and Middle Precambrian orogenesis. Whether this instability is a cause or an effect of the high heat flow is not known. However, it is possible that both are manifestations of subcrustal processes localized along a major structural discontinuity in the Canadian Shield.

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METAMORPHISM IN THE LAKE SUPERIOR REGION, U.S.A., AND ITS RELATION TO CRUSTAL EVOLUTION

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Abstract

Precambrian rocks in the Lake Superior region, U.S.A. range in age from 3800 to 1100 Ma. Thus for purposes of discussion, the rock record is divided into five terranes having similar metamorphic attributes.

Terrane I consists of 3600 Ma old quartzofeldspathic and amphibolitic gneiss metamorphosed to the upper amphibolite or lower granulite facies. Although formed mostly before 3000 Ma, this terrane has been reactivated and metamorphosed a number of times. Terrane II is a typical Archean greenstone-granite terrane characterized by supracrustal rocks generally metamorphosed to the lower greenschist facies. Exceptions are contact metamorphic aureoles near granite plutons and areally extensive, faulted-bounded gneiss belts where metasomatized, amphibolite-facies rocks are common. Stratified rocks deposited 2000-1900 Ma ago in an intracratonic basin centred approximately over the Terrane I — Terrane II boundary comprise Terrane III. Strata overlying Terrane II are metamorphosed to the greenschist facies, whereas those overlying Terrane I are metamorphosed to the upper amphibolite facies, particularly around the flanks of basement highs. Terrane IV, deposited 1800-1500 Ma ago on a gneissic basement, consists predominantly of quartz-rich sedimentary rocks metamorphosed locally to the greenschist facies. Mafic plutonic and volcanic rocks and derivative red beds formed in an incipient rift system at about 1100 Ma ago comprise Terrane V. The stratified rocks are regionally metamorphosed from the zeolite facies to the lower part of greenschist facies whereas the plutonic rocks have contact metamorphic aureoles assignable to the pyroxene- or hornblende-hornfels facies.

Terranes I and II are fundamental crustal segments formed in Archean time. During Proterozoic time, Terranes III and IV represented episodes of subsidence and sedimentation on a craton provided by the older terranes, whereas space for Terrane V was made by the rifting of the earlier-formed terranes. Therefore since the advent of Proterozoic time, the Lake Superior region has been characterized by vertical rather than lateral accretionary processes.

Résumé

L'âge des roches précambriennes de la région du lac Supérieur, aux Etats-Unis, se situe entre 3 800 et 1 100 Ma. Pour cette étude, on a subdivisé la colonne stratigraphique en 5 unités qui présentent des caractères métamorphiques similaires.

Le terrain I consiste en gneiss quartzo-feldspathiques et amphibolitiques, âgés de 3 600 Ma, qui ont été métamorphisés jusqu'au degré du faciès amphibolite supérieure ou du faciès granulite inférieure. Bien qu'il se soit en majeure partie formé il y a plus de 3 000 Ma, le terrain en question a été réactivé et métamorphisé un certain nombre de fois. Le terrain II est un terrain archéen typique, composé de granites et roches vertes, caractérisé par la présence de roches supracrustales généralement métamorphisées dans le faciès schistes verts inférieurs. Il existe cependant des auréoles de contact à proximité des plutons granitiques et des vastes zones gneissiques limitées par des failles, où les roches métasomatisées du faciès amphibolite sont fréquentes. Les roches stratifiées déposées il y a 2 000 ou 1 900 Ma dans un bassin intracratonique, approximativement centré sur la lisière du terrain I et du terrain II, englobent le terrain III. Les strates susjacentes au terrain II sont métamorphisées dans le faciès schistes verts, tandis que celles sus-jacentes au terrain I sont métamorphisées dans le faciès amphibolite supérieure, en particulier sur les versants des surélévations du soubassement. Le terrain IV dont les couches se sont déposées il y a 1 800 à 1 500 Ma sur un soubassement gneissique, est surtout constitué de roches sédimentaires riches en quartz et localement métamorphisées dans le faciès schistes verts. Les roches mafiques, plutoniques et volcaniques, et les "red beds" formés dans un système de rifts en formation, il y a environ 1 100 Ma, constituent le terrain V. Les roches stratifiées ont été soumises au métamorphisme régional et sont passées du faciès à zéolites à la partie la plus basse du faciès schistes verts, tandis qu'au voisinage des roches plutoniques se trouvent des auréoles de contact se situant dans le faciès des pyroxènes ou celui des cornéennes à hornblende.

Les terrains I et II représentent des fragments de la croûte initiale formée pendant l'Archéen. Les terrains III et IV correspondent aux épisodes protérozoïques de subsidence et de sédimentation dans un craton formé par les terrains plus anciens, tandis que l'espace occupé par le terrain V a été créé par fracturation (avec formation de rifts) des terrains formés antérieurement. Par conséquent, depuis le début du Protérozoïque, la région du lac Supérieur a été caractérisée par des processus d'accrétion verticale plutôt que latérale.

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INTRODUCTION

Precambrian rocks in the Lake Superior region underlie parts of Minnesota, northern Wisconsin, and northern Michigan very near the geographic centre of the North American continent (Fig. 1). In terms of surface exposures, the region lies at the southern extremity of the Canadian Shield, but if the subsurface geology is considered, it is located near the centre of the known Precambrian basement of the North American Craton. Except in the north, the Precambrian rocks are overlain by strata of Paleozoic and Mesozoic age which form a thin cover of relatively undisturbed rocks that generally thicken southward and westward. However Precambrian rocks in southwestern Minnesota and south-central Wisconsin are exposed as outliers surrounded by Phanerozoic strata.

Traditionally, geologic studies in the Lake Superior region have focused on the economically important iron- and copper-mining districts. Although these studies have yielded a considerable amount of detailed data, the geologic history of the region as an entity is poorly understood mainly because of a wide diversity of superposed rock types and metamorphic events, ranging in age from about 3800 Ma to less than 1100 Ma before present. Because the rock record is complex and spans a long period of time, an attempt will be made here to describe the metamorphic history of the various rock units, particularly in terms of the crustal evolution of this part of the Canadian Shield.

GENERAL PROCEDURES

The metamorphic map of the Lake Superior region was constructed in three steps. Inasmuch as regional geologic maps are not readily available, it was necessary as a first step to compile a generalized bedrock geologic map (Fig. 2) utilizing a number of published and unpublished sources. In particular, the geology of Minnesota was modified from Morey (1976a), which in turn was generalized from Sloan and Austin (1966), Sims et al. (1970), Austin et al. (1970), Sims and Ojakangas (1972), Southwick and Ojakangas (1972), Morey (1975), Morey and Olsen (1976), Weiblen and Morey (1976), and Sims (1970). Geologic relationships in northern Michigan were compiled from Meshref and Hinze (1970), augmented by recent studies by Hubbard (1975a) and Schmidt (1976). The geology of Wisconsin is based primarily on the provisional map of Sims (1976a) with some additional modifications based on data contained in Dutton and Bradley (1970), LaBerge (1976), Hubbard (1975b), May (1976), and Sims and Peterman (1976).

Figure 1. General map of the North American continent showing the location of the Lake Superior region, U.S.A. relative to the Canadian Shield and to known or inferred Precambrian basement rocks of the North American Craton.

The geologic map departs from conventional stratigraphic usage in the Lake Superior region in that the various rock units are assigned ages according to the classification scheme of the Geological Survey of Canada (Stockwell, 1964) rather than according to the more commonly accepted three-fold subdivision — lower, middle and upper Precambrian — of Goldich et al. (1961). Although this was done strictly for the convenience of the Canadian reader, the two schemes can be used more or less interchangeably with each other and with the W, X, Y, Z scheme of the U.S. Geological Survey (James, 1972). The lower Precambrian (or Precambrian W) is equivalent to the Archean, whereas the middle Precambrian (or Precambrian X) is equivalent to the Aphebian of Canadian usage. The upper Precambrian in the Lake Superior region corresponds to the Paleohelikian and Neohelikian of Canadian usage and Precambrian Y and Z of the U.S. Geological Survey.

The bedrock geologic map was used to plot available data regarding metamorphic minerals and mineral assemblages. Unfortunately, the distribution and amount of data are quite variable from place to place, in part because much of the region is covered by a thick mantle of Quaternary materials. Similarly, the quality of the data, insofar as metamorphic attributes are concerned, is variable. In general, information regarding the higher grade metamorphic rocks (lower amphibolite facies and higher) is good. Commonly these rocks are recognizable in the field as metamorphic rocks, have been reported as such, and most information about them pertains directly to the metamorphism. In contrast, the metamorphic attributes of only slightly metamorphosed rocks (greenschist facies and lower) are poorly known, mainly because the primary nature of these rocks is the feature most commonly described; secondary minerals of metamorphic origin are cursorily treated. A second problem in regard to the quality of data stems from the many localities where other lines of evidence (such as detailed radiometric studies) indicate the presence of superposed metamorphic events. Because the metamorphic character of many of these rocks has not been comprehensively evaluated in the available descriptions, it is not always possible to distinguish primary versus secondary metamorphic mineral assemblages or to correlate various assemblages with specific metamorphic events.

Lastly, the site-specific metamorphic data were interpreted in terms of the bedrock geology to prepare the final metamorphic map (Fig. 3), which utilizes metamorphic units comparable to those used in Geological Survey of Canada Project 740017—Metamorphism in the Canadian Shield. Because the basic data points are so scattered, many of the facies boundaries on the metamorphic map are conjectural. Nonetheless it seems more valuable to generalize the limited data and present an interpretative map than to give a "patchwork" map on which data are so scattered that metamorphic patterns are not apparent.

Table 1

Summary of Precambrian geological events that formed or affected the fundamental basement terranes of Minnesota (modified from Morey, 1976b). Primary events or processes are capitalized. Superposed or subsequent events are listed in lower case. See text for discussion.

Ma	FUNDAMENTAL TERRANES		
	I - Gneiss Terrane	II - Greenstone - Granite Terrane	V - Mid continent rift system
600			
1000	Rifting and low-grade metamorphism	Rifting and contact metamorphism	VOLCANISM, MAFIC PLUTONISM AND CLASTIC SEDIMENTATION
1400	EMPLACEMENT ALKALIC ROCKS metamorphism DEPOSITION TERRANE IV ROCKS FELSIC VOLCANISM AND PLUTONISM	metamorphism ? Dykes	
1800	EMPLACEMENT CALC-ALKALINE ROCKS DEPOSITION TERRANE III ROCKS Dykes metamorphism	DEPOSITION TERRANE III ROCKS Dykes	
2200	Dykes	Dykes	
2600	EMPLACEMENT GRANODIORITE AND QUARTZ MONZONITE (LOW-GRADE METAMORPHISM)	CATACLASIS AND (LOW-GRADE METAMORPHISM) VOLCANISM, SEDIMENTATION AND EMPLACEMENT GRANITOID ROCKS	
3000	EMPLACEMENT HIGH-POTASH GRANITE (deformation and metamorphism ?) Dykes and sills		
3400	TRONDHJEMITIC, TONALITIC AND GRANODIORITIC MAGMA		
3800	Sedimentation, volcanism, etc. ? ? ? ? ?		

fundamental blocks. I suggest here that these same terrane units, with slight modification, can be recognized throughout the Lake Superior region (Table 1).

Several of the terranes defined in Table 1 are lithostratigraphic entities whose boundaries correspond to temporal boundaries as defined or assumed in conventional time-stratigraphic classification schemes. However other terranes are not simple litho- or time-stratigraphic entities. Rather they are the end products of a number of separate geologic events, which did not operate everywhere in the region with equal intensity. Thus each terrane is a geographic entity characterized by distinctly different rock assemblages, tectonic styles, and metamorphic grades.

The use in this report of terrane units having geographic rather than temporal boundaries does not negate or replace in any way the importance of classifying rocks and events in terms of time. The use of these terrane units in preference to time-stratigraphic units only emphasizes the fact that the following discussion is on geologic processes as recorded in the rocks, rather than on the time of these processes **per se.** However, time-stratigraphic relationships remain the primary basis of classification within each terrane, and time is the major basis for relating events in one terrane with those in another terrane.

DISCUSSION OF METAMORPHIC ATTRIBUTES

Conventional analytical procedures would dictate that the various metamorphic events shown on Figure 3 be described sequentially in terms of some time-stratigraphic classification scheme, i.e. Archean metamorphism, Aphebian metamorphism, etc. This procedure has been used in the explanation accompanying the metamorphic map. I suggest, however, that a consideration of metamorphic events in the Lake Superior region in this manner tends to obscure rather than clarify what appear to be some very fundamental relationships in terms of crustal evolution. Thus for purposes of discussion, I propose that the Precambrian rock record be divided into a number of "terrane units", each consisting of a group or series of rocks typified by similar lithic, tectonic, and most importantly, metamorphic attributes.

The concept of terrane units in the Lake Superior region was first used by Morey and Sims (1976) who recognized two fundamentally different basement (Archean) terranes — an older gneiss terrane (3000 Ma or older) and a younger greenstone-granite terrane (2750-2700 Ma in age). Sims (1976a) later expanded on this idea and emphasized the contrasting roles these terranes had on the tectonic evolution of Aphebian strata in the Lake Superior region. Walton et al. (1975) and Morey (1976b) recognized that Precambrian rocks in Minnesota could be divided into five basic terrane units, three of which are fundamental blocks or plates separated by major crustal sutures or rifts. The remaining two terranes are essentially supracratonal features formed on the more

Terrane I — Gneiss Terrane

Rocks assigned to the gneiss terrane are inferred to underlie much of the southern part of the Lake Superior region (Fig. 4). The gneiss terrane is separated from the younger greenstone-granite terrane to the north by a major structural discontinuity that trends diagonally across the central part of Minnesota from approximately 45°30'N to near the west end of Lake Superior. Near Lake Superior, the boundary is transected and displaced southeastward for a distance of about 160 km by a transform fault active during the evolution of Terrane V rocks. In Wisconsin the boundary lies a short distance south of the Gogebic iron range and it is tentatively interpreted as passing eastward through the Marquette trough (Sims, 1976a). The boundary is interpreted as a tectonic surface (Morey and Sims, 1976) and at most places may be a high-angle fault or more likely, a fault zone.

Gneissic terrane rocks are exposed in the Minnesota River valley of southwestern Minnesota (Grant, 1972), in central Minnesota (Morey and Sims, 1976; Morey, in prep.), in the Upper Peninsula of Michigan (James et al., 1961; Cannon and Simmons, 1973; Sims and Peterman, 1976), and in the northeastern part of Wisconsin (Sims, 1976a). In addition the gneissic terrane appears to be sporadically exposed in parts of central Wisconsin (Van Schmus and Anderson, 1977).

In general, the gneiss terrane is best known in the Minnesota River valley where several types of migmatitic gneisses — including hornblende-pyroxene gneiss, garnet-biotite gneiss, tonalitic and granodioritic gneiss, biotite gneiss — and related hybrid rocks have been recognized.

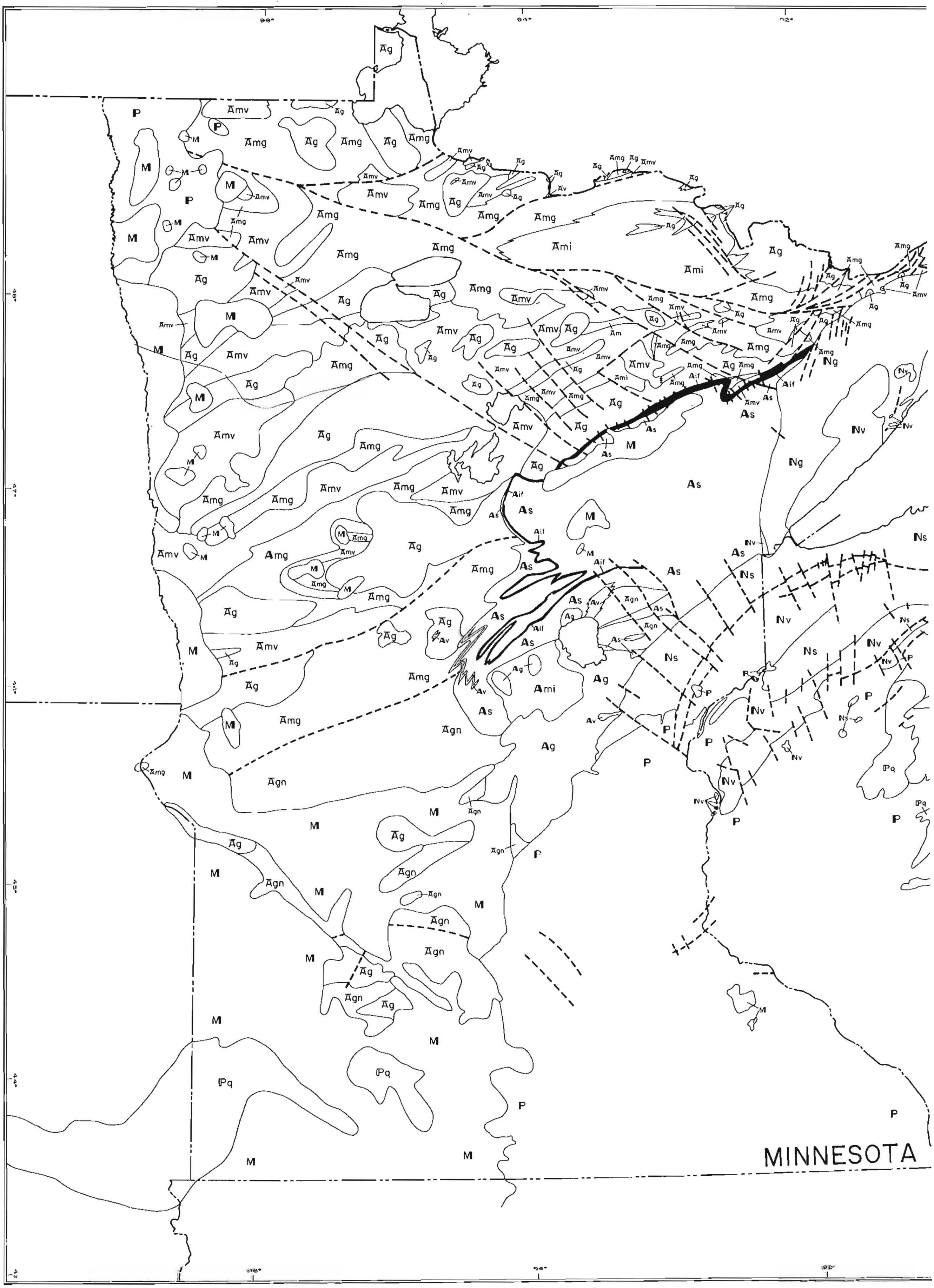

 Figure 2. Bedrock geologic map of the Lake Superior region. See text for summary of sources and discussion.

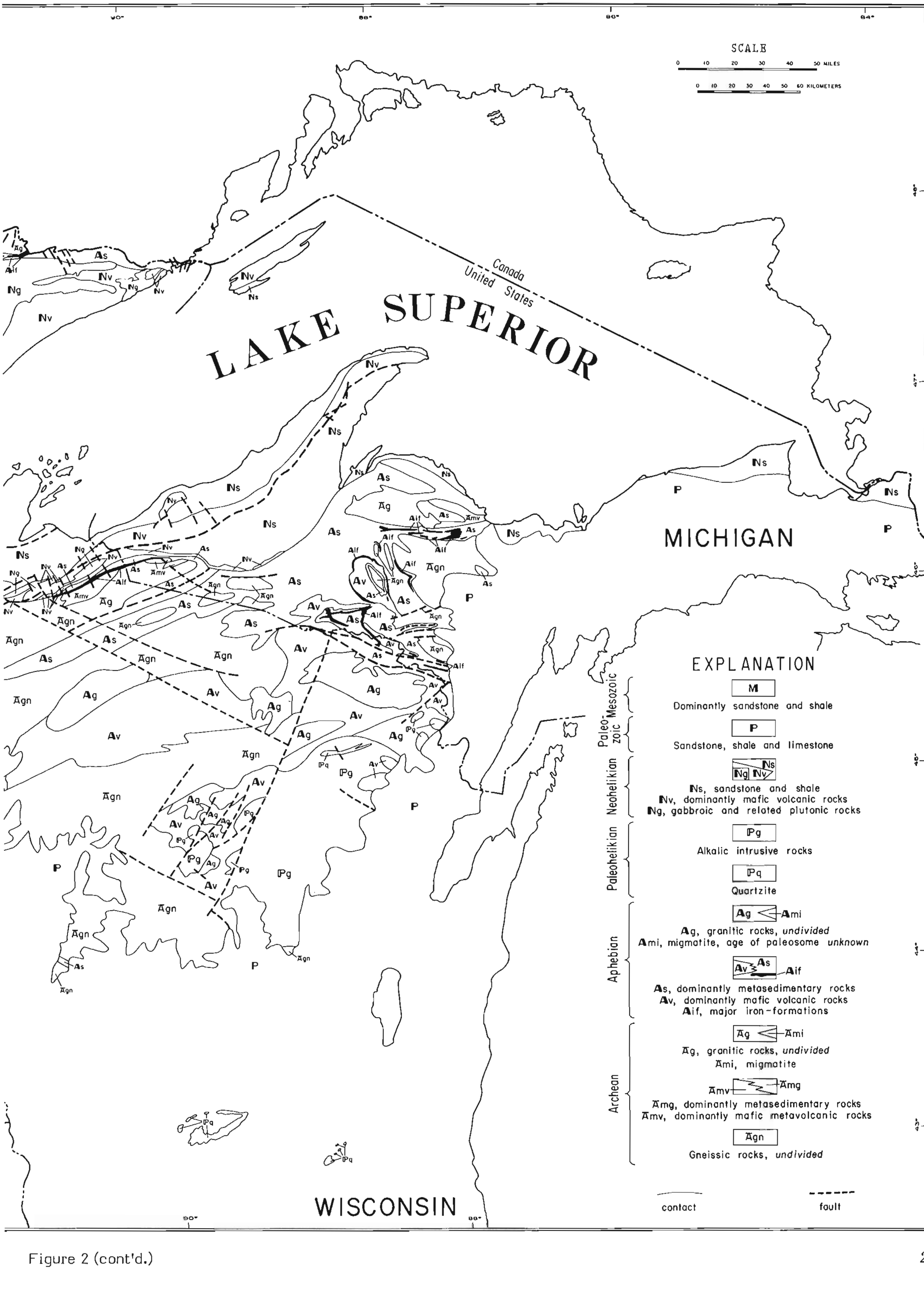

Figure 2 (cont'd.)

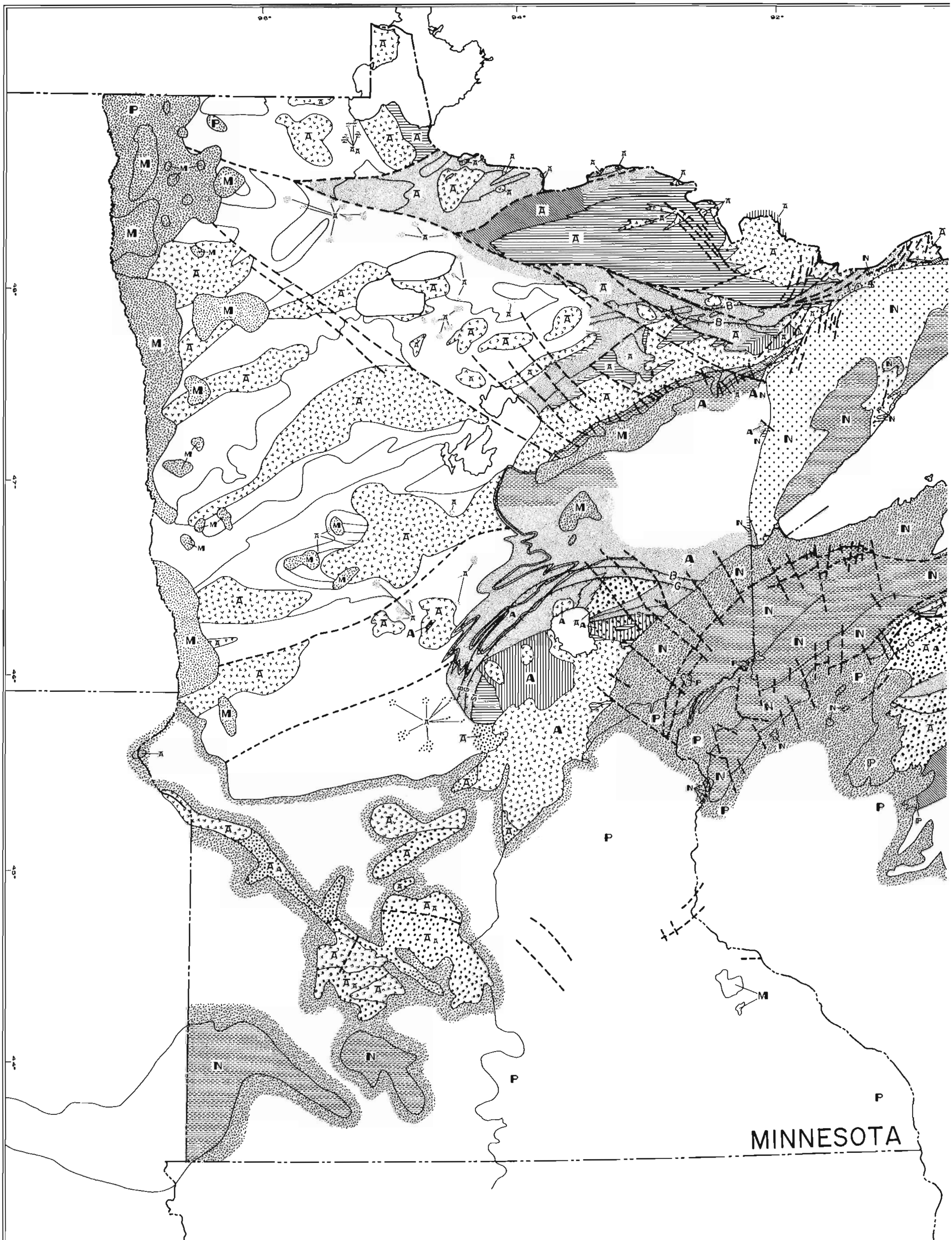

 Figure 3. Map of metamorphic facies in the Lake Superior region. See Tables 2-6 and Figures 4-8 for sources of data.

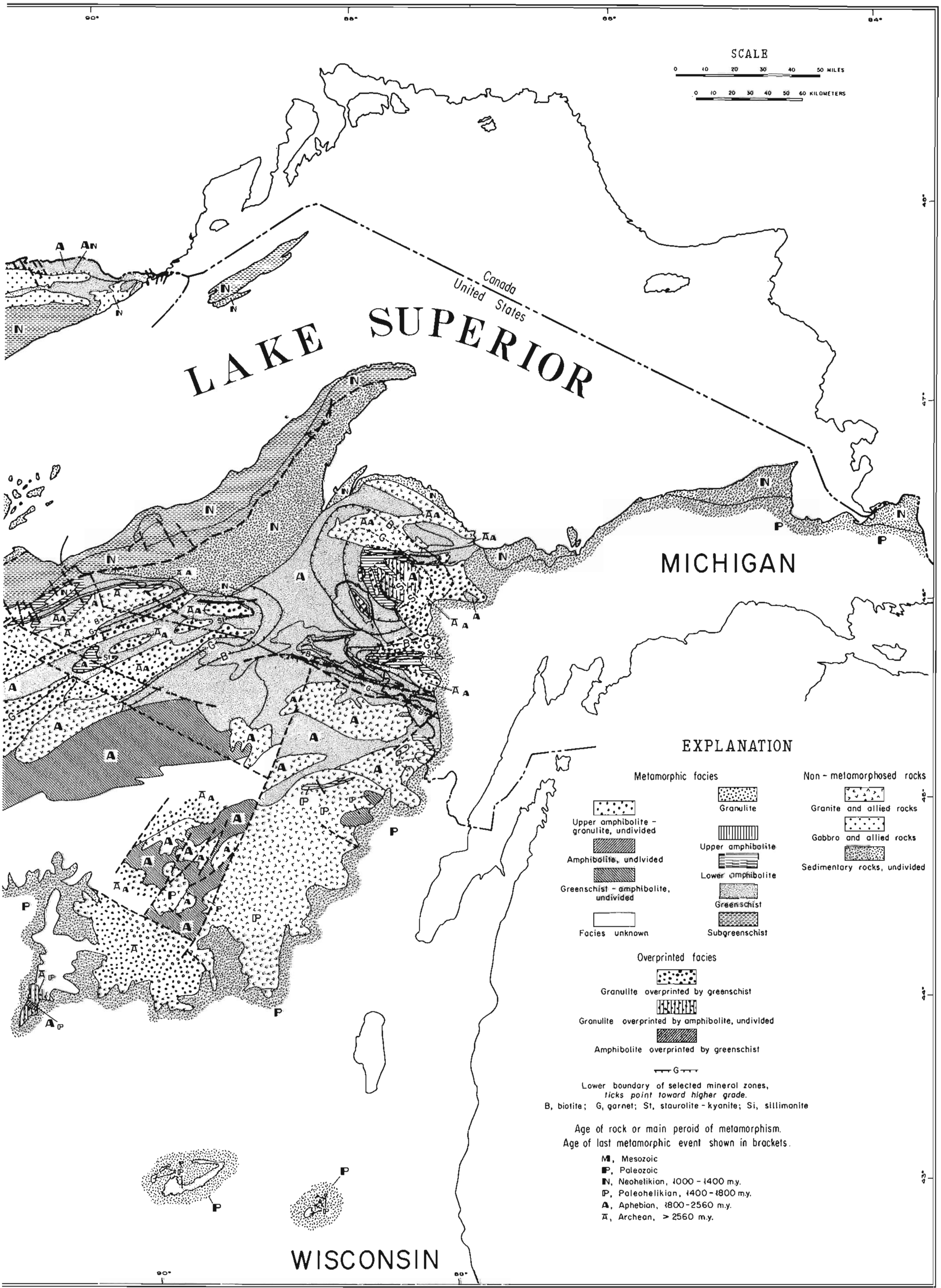

Figure 3 (cont'd.)

Because of a pervasive metamorphic overprint, the original nature of the principal gneissic units is conjectural. The garnet-biotite and biotite gneiss are probably derived from a sedimentary protolith, the hornblende-pyroxene gneiss from a mafic protolith, and the tonalitic and granodioritic gneiss from some felsic protolith.

All gneissic rocks in the Minnesota River valley have been subjected to several periods of folding (Bauer, 1976) and metamorphism that have led to the formation of a completely tectonized fabric characterized by various kinds of mineralogic layering. Thus although a number of lithic units have been recognized (Grant, 1972), they do not necessarily imply a stratigraphic succession, but may merely reflect a repetition of rock types due to structural processes. Mapping of the various lithic units has led to the recognition of a number of moderately open antiforms and synforms that plunge gently east and have axial planes dipping steeply south. However Bauer (1976) has delineated two periods of folding that are younger than the major antiformal structures and one period of folding that may be older. The earlier folding is represented by small, isoclinal, recumbent folds that are coaxial with the major antiforms and synforms. These folds may be remnants of a very early period of folding, or they may have formed during the development of the major structures. Folds younger than the major antiforms and synforms include small to large parasitic folds; their orientations are generally consistent with those of the major antiforms. Still younger minor folds whose axial planes trend in a northwesterly direction and are inclined to the northeast also are present. The latter folds may have formed well after the main period of folding.

Some of the gneiss in the Minnesota River Valley has been dated by Rb-Sr and U-Pb methods at 3600 to 3800 Ma (Hedge and Goldich, 1976). Granitic magma was intruded in the form of small irregular masses, sheets, and **lit-par-lit** injections approximately 3000 Ma ago. Upper amphibolite to lower granulite facies metamorphism affected both the older gneissic and the granitic phases at approximately 3000 Ma, or during the interval from 3000 to 2600 Ma (Goldich et al., 1976). Subsequently the gneiss was invaded by granitic plutons having ages of about 2600 Ma (Goldich et al., 1970; Farhat, 1975). Some retrograde metamorphism of the gneiss terrane appears to have accompanied this period of igneous activity (Wilson, 1976).

Elsewhere in the Lake Superior region, gneissic rocks assigned to Terrane I may be comparable in age to those in the Minnesota River valley, but samples from only a few localities in northern Michigan and central Wisconsin have yielded old ages (Peterman et al., 1976; Van Schmus and Anderson, 1977). Radiometric dating is extremely difficult because the gneiss have been subjected to several different thermal, metamorphic, and tectonic events. For example, in the Minnesota River valley, dykes and stocks of felsic to mafic composition were emplaced at about 1800 Ma. Although in the valley, in east-central Minnesota, these younger granitic intrusions are small and sporadically distributed, temporally equivalent rocks comprise major plutons of batholithic dimensions. The plutonism was accompanied by pervasive metamorphism which affected large segments of the gneiss terrane. Granitic rocks of calc-alkaline affinity having 1900 Ma (Aphebian) ages occur locally in the gneiss terrane of Wisconsin (Van Schmus et al., 1975a). The gneissic terrane also appears to have been affected by a period of felsic volcanism and plutonism at around 1800 Ma (Van Schmus et al., 1975a) and by the emplacement of 1500 Ma old (Van Schmus et al., 1975b) plutonic rocks of alkalic affinity.

The pre-2600 Ma metamorphic history of the gneiss has been studied extensively only in the Minnesota River valley (Table 2 and Fig. 4). In general, metamorphic mineral assemblages so far recognized (Table 2, localities 1-3) indicate both upper amphibolite and granulite facies conditions. However, Himmelberg and Phinney (1967) found no geographic distribution of assemblages indicative of metamorphic zoning, nor could they identify any mineral isograds. Thus they concluded that the various assemblages were isofacial and that the intermingling of pyroxene-rich and hornblende-rich layers was controlled chemically, rather than by fluctuations in temperature or pressure. The temperatures at which the various prograde assemblages formed have not been determined, but Himmelberg (1968) has pointed out that they must have been low enough to allow the hydrous phases to crystallize and that the anhydrous assemblages of similar bulk compositions probably represent beds locally impoverished in water. Although no earlier metamorphic events have been recognized, Grant (1972) argued that the low P_{H_2O} apparently required for granulite-facies metamorphism to occur would be most readily attained either after multiple metamorphic events or after metamorphism of long duration.

Primary mineral assemblages from east-central Minnesota (Table 2, localities 4, 5, and 7) are similar to those from the Minnesota River Valley and indicate upper amphibolite to granulite facies metamorphism. Descriptions of quartzofeldspathic and amphibolitic rocks from elsewhere in the Lake Superior region however, are not sufficiently detailed to precisely determine the metamorphic grade. Exceptions are localities 15-20 (Table 2) from central Wisconsin where prograde assemblages compatible with those from Minnesota have been reported.

There is also considerable evidence of retrograde metamorphism in the gneissic rocks of this terrane. For example, in the Minnesota River valley orthopyroxene is commonly replaced by cummingtonite and iron oxides (Himmelberg and Phinney, 1967). Similarly, primary hornblende and clinopyroxene are replaced by actinolite which in turn has been converted to hornblende. The hornblende to actinolite and actinolite to hornblende reactions suggest a period of low-temperature re-equilibration followed by a period of higher temperature re-equilibration. Other retrograde reactions are indicated by the presence of small biotite flakes around the periphery of primary magnetite grains, and the local alteration of garnet and biotite to chlorite (Himmelberg and Phinney, 1967). Rocks containing potassium feldspar and sillimanite (Table 2, locality 2) are characterized by secondary muscovite surrounding sillimanite and intergrown with myrmekite. This assemblage implies retrogression across the second sillimanite isograd (Grant, 1972). Cryptic zoning in garnet in assemblages containing quartz-plagioclase-biotite-cordierite-garnet-anthophyllite and ilmenite (Grant and Weiblen, 1971) also suggests retrogression in the vicinity of the second sillimanite isograd. Although some of the retrograde reactions noted above may have resulted from cooling after the main period of granulite facies metamorphism, others may have resulted from recrystallization during subsequent periods of thermal metamorphism as reflected in various younger potassium-argon and rubidium-strontium ages.

Many of the gneissic rocks, and particularly those in southwestern Minnesota, east-central Minnesota, and northern Wisconsin, have also been extensively modified by dynamothermal metamorphic processes during cataclasis. In the cataclastic zones, hornblende and pyroxene commonly are replaced by chlorite, calcite, and epidote, the plagioclase is generally saussuritized, and biotite is chloritized. Biotite, potassium feldspar, and hornblende occur as new minerals within many of the cataclastic zones. Although mineral assemblages within the cataclastic zones are generally similar from place to place, dynamothermal metamorphism

Table 2

Inventory of diagnostic metamorphic minerals and mineral assemblages in Terrane I — Gneiss Terrane

Locality	Rock Type	Minerals and/or Mineral Assemblage	Remarks	Reference
1	Hornblende-pyroxene gneiss	QZ-PL bearing		Himmelberg and Phinney, 1967
		HO-OP-CX-MG-IL-(BO)	common	
		HO-OP-CX-BO-MG-IL-(KF)		
		HO-OP-MG-IL-(GA-BO)		
		HO-CP-GA-MG-IL-(KF)		
		HO-SH-HE		
		OP-CX-GA-MG-IL-(HO)		
		OP-CX-MG-IL-(BO)		
		OP-BO-MG-IL		
		OP-MG-IL-(KF)		
		QZ-free, PL-bearing		
		HO-OP-CX-MG-IL-(BO)	common	
		HO-OP-CX-(BO-HE)		
		HO-CX-MG-IL-(BO)		
		HO-MG-IL		
		OP-CX-MG-IL		
		CX-MG-IL		
	Garnet-biotite gneiss	QZ-PL-bearing		
		BO-OP-GA-(KF)	common	
		BO-GA-(KF)	common	
		BO-OP-(KF)		
	Granitic gneiss	QZ-PL-MC-(GA-HE)	common	
		QZ-PL-MC-BI-(GA-HE-RU)	common	
		QZ-MC	rare	
		OP→CU+iron oxides	retrograde	
		HO→AC +HO	retrograde	
		GA+BO→CL		
		CL-EP-AB	cataclastic rocks	
		BO-CL-CA-EP	cataclastic rocks	
2	Granitic Gneiss	QZ-MC-PL-BO		Grant, 1972
	Amphibolite "inclusions"	PL-HO-CX-OP-(QZ-BO)		
	Pelitic gneisses	QZ-PL-HO-GA	CU as descrite grains or homoaxially intergrown with HO	
		PL-HO-CX-SH		
		QZ-PL-BO-CO-GA-AP-IL		
		QZ-KF-PL-BO-CO-SI	AN may be present	
		QZ-KF-PL-BO-GA		
	Amphibolite inclusions	HO-PL-(BO)	BO also occurs as rims on the inclusions	Nielson, 1976
		CX-OP-PL-(BO-MG-IL)	CX rimmed by AC+QZ	
		CX-HO-BO-PL	CX rimmed by AC+QZ	
	Additional assemblages from unspecified rock types.	KF-PL-BO-HO-CX-SH		
		PL-HO-CX-OP		
		QZ-KF-PL-BO-HO-SH		
3	Granitic gneiss	QZ-KF-BO-(HO)		Grant, 1972
	Biotite gneiss	QZ-PL-BO-OP-(MG-GA)		
4	Hypersthene gneiss	QZ-PL-HY-CO-GA-(MC)	primary	Morey, in prep
		HY CU	retrograde	
		CO MU+CL	retrograde	
5	Granitic geniss	QZ-PL-MC-(BO)		Morey and Sims, 1976
	Hornblende-pyroxene gneiss	CP-PL-BO-(HO)		
	Biotite gneiss	PL-QZ-BO-(GA-CO)		
		QZ-MC-GA-CO-(BO)		
6	Granitic gneiss	BO-QZ-PL-MC	primary rock type	Keighin, et. al., 1972
		QZ-PL-MC-MU-BO-EP	cataclastic rocks	Morey and Sims, 1976
7	Hornblende-pyroxene gneiss	CX-HO-BO-PL		Morey and Sims, 1976
8	Granitic gneiss	QZ-MC-PL-(BO-MC-AB)	BO-MC-AB retrograde minerals	Morey and Sims, 1976
9	Granitic gneiss	QZ-MC-PL-(BO-AB)		Morey and Sims, 1976
	Hornblende gneiss	QZ-HO-PL-MC		
10	Granitic gneiss	BO-QZ-PL	primary rock type	Sims and Peterman, 1976
		BO-MC-QZ-(CL)	cataclastic zones	
11	Amphibolite	Ho-Pl-(Bo-Cl-Ep-Mu)	trace minerals secondary components	Gair, 1975
	Granitic gneiss	Pl-Mc-Qz-(Bo-Cl)		
12	Granitic gneiss	Mc-Qz-Pl-(Bo-Mu-Cl)	Bo→Cl	James, et. al., 1961
	Schist	Cl-Qz-Fs-(Ca-Mg-Ep-Sh-Ga)	inclusions in granitic gneiss	
	Amphibolite	Ho-Pl		
		Ho-Pl-Bo		
	Banded gray gneiss	Pl-Kf-Qz-Bo-(Mu-Cl-Ep-Ap)		
13	Gray gneiss	Pl-Qz-Bo(Mu-Mc-Ab)		Bayley et. al., 1966
	Amphibolite	Ho-Pl-(Bo-Qz-Sh-Ep-Ca)		
14	Gneiss, schist, amphibolite and migmatite	Sillimanite grade		Dutton and Bradley, 1970
15	Quartzofeldspathic gneiss	QZ-MC-PL-BO-HO-(EP-SH)		Maas and Medaris, 1976
	amphibolite	HO-PL-SH-(BO-EP)		
16	Migmatitic gneiss	QZ-PL-MC	textural evidence of several periods of metamorphism	Van Schmus and Anderson, 1977
		BO-HO-PL-(CH-EP)		
17	Migmatitic gneiss	QZ-PL-MC	textural evidence of several periods of metamorphism	Van Schmus and Anderson, 1977
		QZ-PL-BO-HO		
18	Amphibolite gneiss	HO-PL-(GA)	HO+EP formed late	Cummings, 1975
	Amphibolite schist	QZ-PL-CU-HO		
19	Granitic gneiss	PL-QZ-HO(GA)		
	Gneissic amphibolite	GA-HO		
		HO-PL-CU		
	Garnetiferous biotite gneiss	QZ-PL-MU-GA		
20	Amphibolite	KY-ST-MU-QZ-BO		Mudrey, written comm.

See Appendix I for explanation of symbols and Figure 4 for localities. Upper case symbols refer to mineral assemblages; lower case symbols to diagnostic minerals.

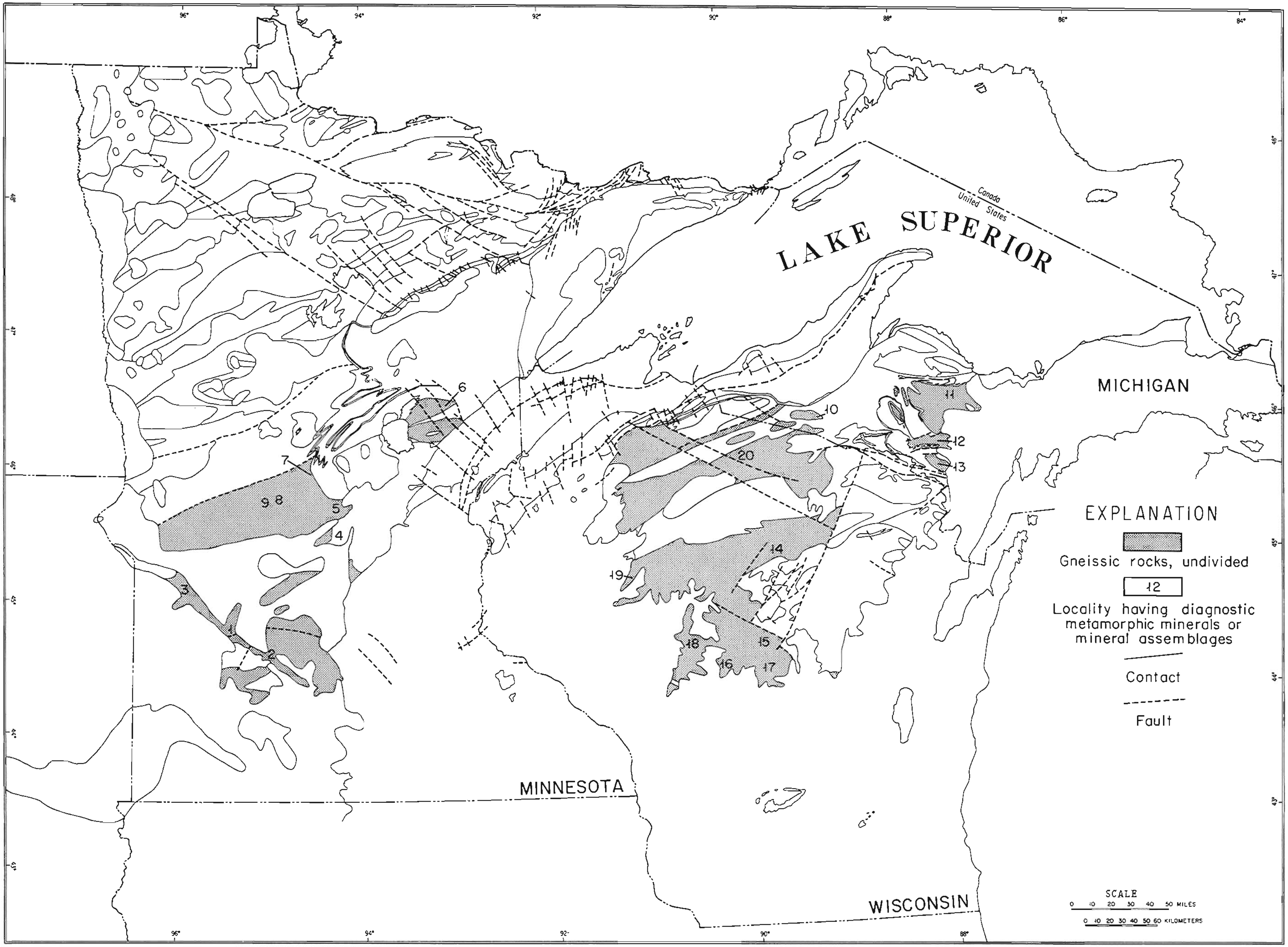

Figure 4. Bedrock geologic map of the Lake Superior region showing the known or inferred distribution of rocks assigned to Terrane I — Gneissic Terrane. See Table 2 for inventory of diagnostic metamorphic minerals and mineral assemblages.

did not occur everywhere at the same time. For example, cataclastic zones in the Minnesota River valley formed between 2000 and 1900 Ma (Hanson and Himmelberg, 1967), whereas those in east-central Minnesota and northern Wisconsin formed around 1800 Ma ago (Keighin et al., 1972; Sims, 1976a).

Terrane II — Greenstone-Granite Terrane

Rocks that characterize the greenstone-granite terrane are exposed over a wide area in northern Minnesota and to a lesser extent in northern Wisconsin and northern Michigan (Fig. 5). In general this terrane, which is typical of the Superior Province of the Canadian Shield, consists of a thick succession of subaqueous volcanic rocks, derivative sedimentary rocks, and intrusive granitic rocks. The volcanic-sedimentary successions constitute complexly folded and faulted, east-trending belts between flanking granitic batholiths. Several periods of folding occurred contemporaneously with the emplacement of the oldest plutonic rocks recognized, which range in composition from tonalite to granite (Sims, 1976b). Metamorphism of the supracrustal rocks to greenschist and locally upper amphibolite facies accompanied the several periods of folding. Later, anorogenic bodies of monzonite, quartz monzonite, and syenite were emplaced in the supracrustal rocks, apparently as diapirs. Extensive faulting accompanied and followed emplacement of the igneous rocks.

Radiometric data from numerous localities indicate that the evolution of this terrane, at least in the Lake Superior region, occurred during a short-lived volcanic-plutonic event about 2750-2700 Ma ago (Jahn et al., 1974). Furthermore, these studies indicate that once formed, the terrane has remained essentially stable and unaffected by younger events except for cataclasis and several periods of dyke emplacement. In addition, parts of the terrane were affected by contact metamorphic processes related to the emplacement of Terrane V rocks.

Currently available data bearing on the metamorphic attributes of the greenstone-granite terrane are summarized in Table 3 and Figure 5. Inasmuch as deformation, metamorphism, and plutonism are closely related, the terrane is arbitrarily subdivided into three litho-tectonic elements, each having characteristic metamorphic attributes. These are: (1) supracrustal units or so-called greenstone belts, (2) batholitic units, and (3) fault-bounded "granite-migmatite massifs".

Supracrustal Units

The supracrustal units consist of several mafic to felsic volcanic sequences complexly intercalated with sedimentary rocks derived from the volcanic rocks. The volcanic sequences are predominantly composed of subaqueous basalt and synvolcanic diabase, but include dacitic to rhyodacitic volcanogenic and volcaniclastic rocks. The sedimentary rocks are mainly greywacke and shale, but beds of banded iron formation, chert, carbonaceous shale, and rare siliceous marble also are present (Morey et al., 1970; Green, 1970). All of these rocks have been metamorphosed, generally to at least the greenschist facies. Quartzofeldspathic rocks contain chlorite, muscovite, albite, quartz, and epidote, whereas the mafic rocks contain chlorite, calcite, tremolite or actinolite, epidote, and quartz. However, incomplete recrystallization is widespread as shown by well-preserved bedding characteristics and primary textures. In addition, zoned plagioclase crystals and relict hornblende grains are widespread in the felsic volcanogenic and volcaniclastic rocks, as are relict augite and labradorite in the mafic rocks.

In general, the metamorphic grade within the greenschist facies increases in intensity toward bounding granite masses or major faults (Fig. 3) so that in some places it is possible to define a biotite isograd. More typically however, mineralogic zoning is not well developed in the supracrustal rocks mainly because of their bulk chemistry. Many of the pelitic rocks inferred to be at or above the biotite isograd lack biotite because they lack K_2O. Moreover those that do contain K_2O are characterized by K-feldspar and chlorite. Similarly other rocks are too rich in Ca and too poor in Al to contain muscovite or other peraluminous silicates.

Retrograde metamorphism of the supracrustal rocks is widespread as indicated by mineralogic transformations such as: (1) the alteration of biotite to chlorite and/or pumpellyite, (2) alteration of plagioclase to epidote, and (3) sericitization of metamorphic plagioclase. Whether these transformations represent re-equilibration in response to declining temperatures during the latter stages of metamorphism at 2700 Ma, or whether they formed in response to some subsequent period of cataclasis or very low grade metamorphism has not been established (Hart and Davis, 1969; Hanson and Malhotra, 1971).

Batholithic units

Batholithic rocks that have intruded the supracrustal rocks form a second litho-tectonic segment in the greenstone-granite terrane. In northern Minnesota where they have been well studied, many of the batholithic units are composite structural entities consisting of a number of smaller plutons. In general these plutons are of variable size and may be divided into older syntectonic units of tonalitic to granitic composition and younger posttectonic rocks of monzonitic to quartz-monzonitic composition.

The syntectonic igneous rocks are characterized by foliated textures, contacts that are conformable with the internal structure of the adjacent supracrustal rocks, abundant inclusions, and wide metamorphic aureoles ranging from the greenschist facies to the upper amphibolite facies; the inner parts of some of the aureoles are migmatized. All of the contact metamorphosed supracrustal rocks at localities 30-32 (Table 3), for example, are above the biotite isograd. The metamorphic grade increases toward the batholithic rocks and is characterized by: (1) the progressive loss of relict volcanic and clastic textures, (2) the progressive disruption of layering, and (3) a progressive increase in the degree of migmatization (Griffin, 1967; Griffin and Morey, 1969). In the outer part of the aureole, mafic rocks contain plagioclase more calcic than An_{25}, accompanied by hornblende and epidote. In the inner part, epidote and calcite disappear from the equilibrium assemblages, but biotite, hornblende, and sodic plagioclase persist. Each of these minerals displays a wide range of solid solution compositions. Consequently changes in pressure, temperature, and whole-rock bulk composition tend to be reflected as variations in the composition and relative abundance of the phases, rather than by the formation of new minerals. For this reason, it has been impossible to map mineral isograds, despite the textural evidence for a steep metamorphic gradient.

Metasomatic activity in the inner parts of some of the contact metamorphic aureoles is characterized mostly by the addition of K_2O leading to the formation of antiperthites and to the alteration of hornblende to biotite and epidote. In some of the high grade amphibolites, retrograde reactions include the partial transformation of hornblende to vermicular epidote and cummingtonite. Other retrograde reactions include: (1) the chloritization of biotite, (2) the breakdown of hornblende to chlorite and sphene, (3) the replacement of plagioclase by sericite and/or epidote, and (4) the breakdown of biotite to low Ab-K-feldspar and

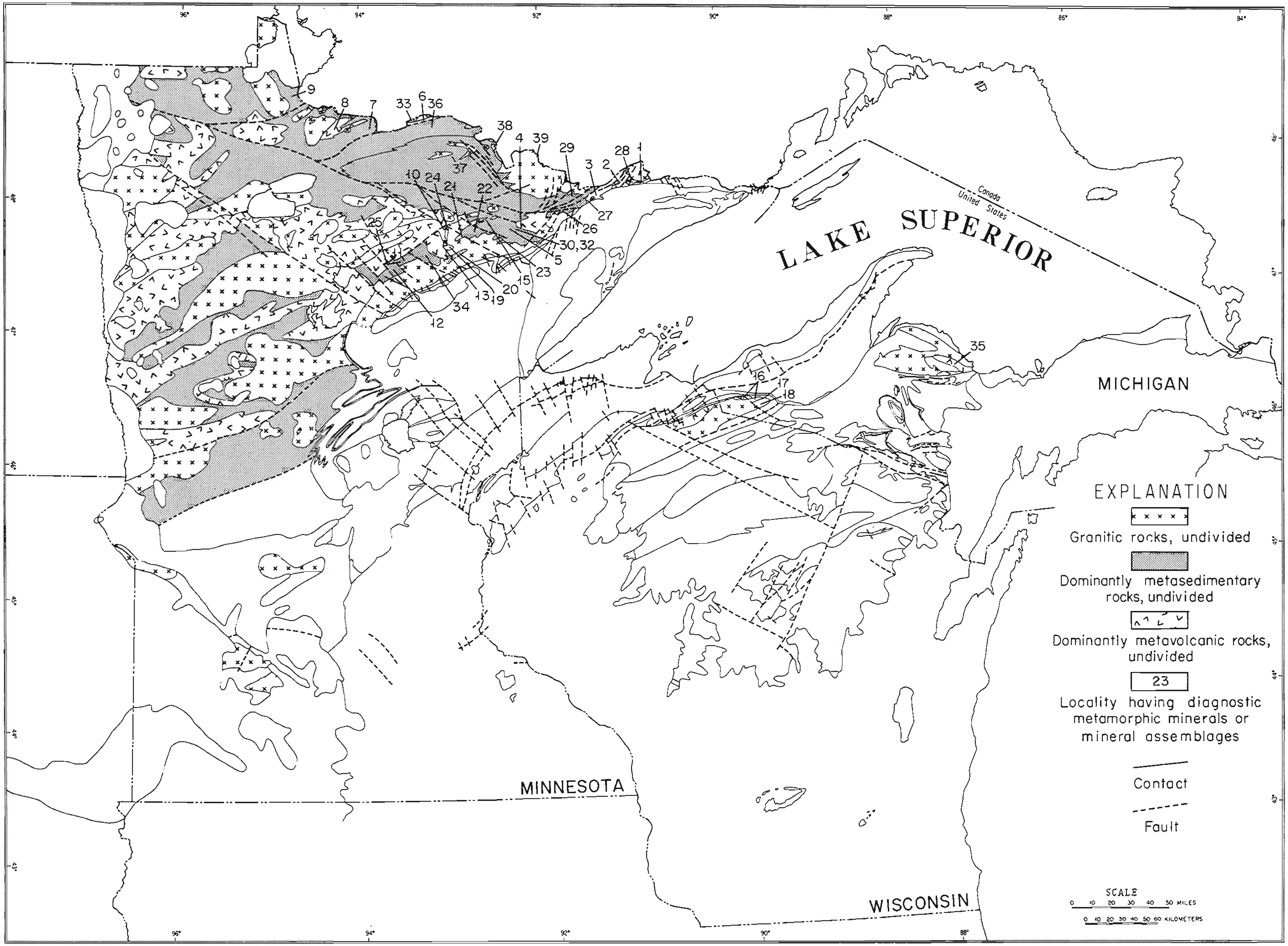

Figure 5. Bedrock geologic map of the Lake Superior region showing the distribution of rocks assigned to Terrane II — Greenstone-granite Terrane. See Table 3 for inventory of diagnostic metamorphic minerals and mineral assemblages.

magnetite in antiperthite-bearing rocks. Griffin and Morey (1969) suggested the minimum temperature for the latter reaction at P_{Total} equals 2 kb[1] is 520°C. Thus the maximum temperature for the prograde metamorphic reactions must have been somewhat higher.

In contrast to the syntectonic granitoid rocks, the posttectonic rocks of generally quartz-monzonitic composition truncate the regional structures and have steep primary foliations that are discordant with the regional structure, but tend to be concordant with the structures in immediately adjacent wall rocks. In addition to the major batholithic units, posttectonic rocks of generally syenitic composition form small plutons that are widely scattered along the supracrustal belts. Regardless of their composition or stratigraphic setting, these posttectonic rocks have narrow metamorphic aureoles characterized by a lack of primary textures except for layering. The maximum grade obtained is generally of the lower amphibolite facies, characterized by mineral assemblages that contain garnet, hornblende, plagioclase, sphene, and diopside in addition to epidote, zoisite, and actinolite.

Granite-migmatite massifs

Fault-bounded, linear belts characterized by large quantities of biotite schist and amphibolite intercalated with igneous rocks of dioritic to granitic composition comprise a third discrete litho-tectonic segment in the greenstone-granite terrane. In general, these rocks have been metamorphosed to upper amphibolite facies, have had a complex history of injection, anatexis, and metasomatism, and are associated with syntectonic igneous rocks. Therefore they are analogous in many ways to the contact metamorphosed supracrustal rocks associated with the batholithic units. However the fact that these fault-bounded blocks are areally extensive and contain much more migmatitic material than do the other batholithic units in northern Minnesota led Southwick (1972) to refer to them as "granite-migmatite" massifs.

Supracrustal rocks along the outer fringes of the massif in northern Minnesota (Table 3, locality 36) have metamorphic mineral assemblages similar to those in adjacent supracrustal rocks. However, primary textures within the massif have been totally obliterated except for layering, and the fabric is entirely metamorphic in origin. Moreover primary layering is recognizable only by contrasting differences in grain sizes and differing modal proportions of plagioclase, quartz, biotite, and muscovite. Sill-like bodies of igneous material, ranging in thickness from a few centimetres to tens or hundreds of metres, become more abundant in the interior portions of the massif (Table 3, locality 37). In places the supracrustal rocks contain garnet, staurolite, hornblende, and epidote, but the areal distribution of these minerals appears to be controlled by original bulk compositional differences rather than by differences in temperature and/or pressure. These rocks in turn pass into a stromatoform migmatite characterized by roughly equal proportions of schist and granite; much of the schist contains so many quartzofeldspathic stringers that the combined rock appears gneissic. The core of the massif consists of granite containing numerous biotite-rich inclusions that show all stages of transformation to rocks having the texture and composition of the enclosing granite. These inclusion-rich zones grade transitionally into areas consisting almost entirely of monotonous granite whose uniformity is interpreted only by scattered zones of nebulitic inclusions.

Supracrustal rocks within several hundred metres of granite bodies in the core of the massif (Table 3, localities 38 and 39) contain various combinations of sillimanite, cordierite, staurolite, and garnet along with biotite, oligoclase and quartz. There is excellent textural evidence that the reaction garnet + staurolite + quartz → cordierite has taken place. These assemblages suggest maximum temperatures and pressures in the neighborhood of 600-700°C and 3-5 kb (Southwick, 1976), a metamorphic environment compatible with inferred temperatures and pressures at the time of crystallization of the granitic phases in the massif (Southwick, 1972).

In summary, metamorphism and tectonism in the greenstone-granite terrane were closely related processes. Although precise temporal relationships between the regional metamorphism and the contact metamorphism associated with syntectonic igneous rocks are not well understood, it seems likely that the former is a more extensively developed part of the latter and that both occurred as the igneous rocks were emplaced under conditions characterized by steep thermal gradients (Sims, 1976b). Moreover it also seems likely that temperatures of 600-700°C and pressures of 3-5 kb were attained, at least locally, during this metamorphic event. In contrast, strongly deformed wall-rock zones and narrow metamorphic aureoles suggest that the somewhat younger, posttectonic monzonitic magmas were rapidly emplaced to very shallow levels as diapirs (Sims, 1976b). Although the metamorphic history of the granite-migmatite massif in northern Minnesota is grossly similar to that of the contact metamorphosed rocks, the areal extent of the massif suggests that it evolved as an independent entity under tectonic conditions which as yet are not completely understood.

Terrane III — Intracratonal Stratified Rocks

Rocks assignable to this terrane comprise a thick succession of metasedimentary and metavolcanic rocks of Aphebian age which were deposited in an intracratonic basin centred over and approximately parallel with the boundary between Terranes I and II (Fig. 6). In general these rocks record a complete transition in deposition from that of a stable craton to that of a "eugeosynclinal" environment (Bayley and James, 1973; Sims, 1976b). However the nature of the lithic fill varies considerably from place to place, and this variability appears to be related, at least in part, to contrasting kinds of basement rocks. To the north, where the basement in northern Minnesota and adjoining Ontario consists predominantly of Terrane II granitic rocks, sedimentation occurred within a single depositional cycle, starting with well sorted, clastic detritus characteristic of a stable shelf and ending with fine sand and mud characteristic of a deep basin having poor circulation. Volcanic rocks, mostly of mafic composition, are only a minor component in these rocks. In contrast, sedimentation near the Terrane I-Terrane II boundary in east-central Minnesota, northwestern Wisconsin, and adjoining parts of Michigan was more complex and involved at least two depositional cycles separated by an unconformity (Marsden, 1972; Morey, 1977; Aldrich, 1929). The younger cycle is similar to that in northern Minnesota, whereas the older cycle records shallow to deep to shallow sedimentation. Furthermore both cycles are characterized by locally abundant volcanogenic material of both mafic and felsic composition (Morey, 1977; Prinz, 1976).

Sedimentation was even more complex in those parts of northern Michigan that lie well to the south of the Terrane I-Terrane II boundary. At least three depositional cycles have been recognized (Bayley and James, 1973) and the sequence is characterized by abundant subaqueous volcanic rocks of basaltic composition, although intermediate and felsic varieties also are present. Still farther to the south, in northeastern and north-central Wisconsin, Terrane III rocks are predominantly subaqueous mafic and pyroclastic, felsic volcanic rocks. Although information regarding precise stratigraphic relationships is lacking, it is tempting to suggest

[1] 1 kb = 1 x 10^5 kPa

Table 3

Inventory of diagnostic metamorphic minerals and mineral assemblages in Terrane II — Greenstone-granite Terrane.

Locality	Rock Type	Minerals and/or Mineral Assemblage	Remarks	Reference
		I. Regionally metamorphosed supracrustal rocks		
1	Metabasalt	AB-AC-CL-EP-CA-QZ	relict AU and CX AU→AC	Weiblen et. al., 1972
	Felsic volcanogenic and volcaniclastic rocks	QZ-PL-CL-MU	relict textures and detritial HO HO→AC	
2.	Graywacke	CL-CA-EP	in matrix only	McLimmons, 1972
3	Pelitic rocks	CL-MU-QZ-PL-(MC)	incomplete textural reequilibration	Green, 1970
	Mafic rocks	PL-CL-AC-CA	relict AU→AC	
	Iron-formation	QZ-SP-MG-HE		
4	Mafic rocks below BO isograd	CL-CA-EP-TR or AC	relict AU and PL	Sims, 1972
	Mafic rocks above BO isograd	BO-CL-CA-EP-TR or AC		
5	Graywackes etc. below BO isograd	CL-MU-AB-QZ-EP	incomplete textural and mineralogic reequilibration-- relict HO and zoned PL.	Sims, 1972
	Graywackes etc. above BO isograd	BO-CL-MU-AB-QZ-EP		
	iron-formation	QZ-MG or HE QZ-CA-AK QZ-MN-SP-MG		Sims, 1972
	Lamprophyres	CL-EP-CA-(AC)	incomplete textural and mineralogic reequilibration. HO→AC CX→QZ+EP+AC BO→CL+PU	Geldon, 1972
6	Bootite schist	Bo-Qz-Pl-(Ga) Bo-Qz-Ep-Pl-Mg	incomplete textural reequilibration BO→CL	Ojakangas, 1972
	Tuffaceous schists	Bo-Qz;Cl-Qz;Mu-Qz;Qz-Pl-Bo;Cl-Qz-Pl; Cl-Ac-Pl-Qz;Ac-Pl-Qz	Individual laminae BO→CL	
	Metabasalt	Ca-Ab-Cl Ab-Bo-Mu		
	Metasedimentary rocks	CL-BO-GA-CP-(QZ) BO-MU-GA-(QZ)		Hart & Davis, 1969
7	Felsic volcaniclastic rocks	Bo-Ep-Cl-Ac-Ca-Qz-Pl-(Ga)	detritial HO	Ojakangas, et. al., 1977
8	Metabasalt	Cl-Ac-Ep-Pl	detritial HO	Ojakangas, et. al., 1977
9	Amphibolite	Ho-Pl-Ca-(Qz)		Sims and Ojakangas, 1972
10	Metasedimentary rocks	QZ-PL-BO QZ-PL-BO-MC	relict textures	Viswanathan, 1971
11	Metasedimentary rocks	QZ-PL-BO QZ-PL-BO-MC CL-EP-AC-AB-(QZ)	relict textures	Viswanathan, 1971
12	Metasedimentary rocks	QZ-PL-BO QZ-PL-BO-(CL)		Viswanathan, 1971
13	Metabasalt	CL-EP-AC-PL		Viswanathan, 1971
	Metasedimentary rocks	BO-PL-QZ-MC		
14	Metabasalt	CL-EP-AC-AB-(QZ-MC)		Viswanathan, 1971
	Metasedimentary rocks	QZ-PL-BO QZ-PL-MC-BO		
15	Metasedimentary rocks	CL-MU-AB-QZ-EP-BO	relict textures	Viswanathan, 1971
16	Mafic metavolcanic rocks	Cl-Bo-Pl-Qz	relict textures	Schmidt, 1976
		II. Contact metamorphosed supracrustal rocks associated with post-tectonic granitic intrusions		
17	Mafic metavolcanic rocks	Pl-Ho-Qz-Mc	contact metamorphic equivalents of mafic rocks at locality 16	Schmidt, 1976
18	Amphibolitic gneiss	QZ-PL-HO-(BO)	original textures obliterated; high grade equivalent of locality 16 or an older phase	Schmidt, 1976
	Schist	QZ-PL-BO	BO→CL+EP due to superposed Penokean metamorphism	
19	Amphibolite	QZ-PL-BO-HO	high grade equivalent of metabasalt at locality 13	Viswanathan, 1971
20	Amphibolite	HO-PL-(QZ-MC-SH)	recrystallized and foliated	Viswanathan, 1971
21	Biotite schist	QZ-PL-BO-(TR-GA-SH-HO)	recrystallized and foliated	Viswanathan, 1971
22	Stripped amphibolite	HO-PL-EP-SH EP-HO-PL HO-EP-PL-QZ MC-PL-HO	mineral assemblages in individual layers	Viswanathan, 1971
23	Amphibolite	HO-PL-MC-(QZ-EP-SH) HO-PL-(QZ)	PL→EP GA and SH occur locally near granite contact	Viswanathan, 1971
24	Hornblende schist	HO-BO-QZ-PL-MC HO-PL-EP-MC-QZ	high grade equivalent of locality 10	Viswanathan, 1971
25	Biotite schist	BO-HO-QZ-MC-PL	high grade equivalent of locality 11	Viswanathan, 1971

Table 3 (cont'd.)

Locality	Rock Type	Minerals and/or Mineral Assemblage	Remarks	Reference
26	Schist	QZ-PL-BO-HO-EP-MC BO-CL-MG-EP		Green, 1970
	Metabasalt	HO-BO-PL-EP CX-HO-PL-(QZ-EP)		
	Iron-formation	CX-GU-QZ-MG QZ-GU-MG-(BO-MG-SP)		
	Schist	CU-HO-BO-PL-MG CU-BO-GA-CL-QZ-PL		
	Amphibolite	CU-HO-GA-BO-(QZ or PL)		
27	Metasedimentary rocks		CO and BO near contact with granite stock	Green, 1970
28	Amphibolite	HO-PL-QZ	high grade equivalent of locality 2	Weiblen et. al., 1972
29	Metabasalt	PL-EP-AC-CL		Green, 1970
	Metasedimentary rocks	BO-HO-PL-QZ-EP		
	III. Contact metamorphosed supracrustal rocks associated with syntectonic granitic intrusions			
30	Felsic volcanic and clastic rocks, undivided	QZ-free, PL-bearing EP BO-HO HO-EP HO-DI-EP BO-HO-CA BO-HO-AC MC-BO-HO-EP-CA	Relict volcanic and clastic textures preserved; little or no disruption of sedimentary layering PL→MU&/or EP BO→CL→PH HO→CL BO→CL+MC+opaques	Griffin, 1967
		QZ- and PL-bearing BO HO BO-HO BO-EP BO-GA BO-MU HO-EP HO-DI HO-GA DI HO-EP-CA BO-HO-EP BO-HO-CA BO-AC-CA HO-CL-EP BO-HO-CA-EP HO-EP-DO-CA		
31	Felsic volcanic and clastic rocks, undivided	BO HO BO-HO HO-DI MC-BO MC-HO MC-BO-HO	No relict textures preserved; layering generally continuous through locally distorted; no migmatization	Griffin and Morey, 1969
	Metabasalt	HO-PL HO-PL-DI HO-PL-QZ HO-PL-DI-QZ		
32	Felsic volcanic and clastic rocks and metabasalts, undivided	Mineral assemblages similar to those at locality 31	No relict textures preserved; layering discontinuous and lenticular; migmatization and metasomatism common HO+K_2O→BO+EP HO→EP+CU BO→CL HO→CL+SH PL→MU &/or EP	Griffin and Morey, 1969
33	Metabasalt	HO-PL-QZ		Hart and Davis, 1969
	Metasedimentary rocks	BO-MU-MC-CX		
	Paragneiss	QZ-PL-MI		
34	Migmatite	HO-PL AC-PL-HO EP-HO-PL BO-QZ-PL QZ-PL-HO QZ-PL-HO-BO-(EP-SH) HO-DI-SH-PL-QZ	granitic neozome	Viswanathan, 1971
35	Metabasalt and associated rocks	Generally of greenschist facies (maximum garnet zone)	metamorphism in part related to Terrane III	Prinz et. al., 1975
	IV Metasedimentary and associated rocks in granite-migmatite massifs.			
36	Biotite schist	BO-QZ-PL-(GA) BO-QZ-EP-PL-MG	no relict textures preserved BO→CL	Ojakangas, 1972
	Metabasalt	CL-AC-EP-PL	no relict textures preserved	
37	Biotite schist	BO-QZ-PL-(GA)	no relict textures preserved; local metasomatism	Southwick, 1972
	Metabasalt	HO-PL-EP-(QZ)		
38	Paragneiss	SI-CO-ST-GA-SA(?)-(QZ-PL-BO)		Southwick, 1976
39	Paragneiss	GA-ST-CO-QZ-(PL)	GA+ST+QZ→CO	Southwick, pers. comm.

See Appendix I for explanation of symbols and Figure 5 for localities. Upper case symbols refer to mineral assemblages; lower case symbols to diagnostic minerals.

Table 4

Inventory of diagnostic metamorphic minerals and mineral assemblages in Terrane III — Intracratonal supracrustal rocks

Locality	Rock Type	Minerals and/orMineral Assemblage	Remarks	Reference
1	Iron-formation	Qz-Gr-Sp-He-Ca-Do-Ak-(Mn-Ch-Mg-Pi)	"unmetamorphosed"--specific assemblages not identified	Floran and Papike, in press
2	Argillite and graywacke	MU-CL-AB-QZ-(PI) MU-CL-QZ-(PI)	relict textures preserved MU is a $1M_d$ polytype	Morey, 1969
	Concretions in argillite	CA-QZ-AB-MC-MU-CL DO-QZ-PL-MU-CL CA-DO-QZ-PL-MU-CL	QZ &AB detritial	
3	Iron-formation	Qz-Mn-Sp-Ca-Do-Ak-Sd-Mg-(Ch-Gr-He-Pi)	Gr+Qz→Mn Cl+Qz→Sp Sp+H_2O→Mn+Cl+QZ Gr+O_2→Mg+Mn+H_2O Si+O_2→Mg+CO_2 Si+O_2→Mg+Ak+CO_2	French, 1968
	Quartzite	Qz-Ch-(Sp) Qz-Mc-Ch-(Cl)	detritial textures preserved	French, 1968
	Argillite and graywacke	QZ-AB-CL-MU	QZ & AB detritial	
4	Argillite	Qz-Pl-Cl-Mu-(Ca-Do)	detritial textures preserved no slaty cleavage	Marsden, 1972
5	Slate	Qz-Pl-Cl-Mu-(Ca-Do)	well developed slaty cleavage	Marsden, 1972
6	Graywacke and slate	QZ-PL-CL-MU-(CA-DO-PI-AB)	detritial QZ & PL AB occurs as rims on PL	Morey and Ojakangas, 1970
	Concretions in slate	CA-DO-AB-QZ-(MU-CL)		
7	Phyllite and metagraywacke	QZ-PL-MU-(CL-CA-DO-PI)	detritial textures preserved in graywacke PL is albitic	Morey, in prep
	Metadiabase	CL-EP-AC-(AB)	relict PL & AU AU→AC AB occurs as rims on PL	
8	Biotite schist and metagraywacke	BO-MU-QZ-PL-(CL)	PL is oligoclase	Morey, in prep
	Metadiabase	EP-AC-BO-(CL-AB)	relict textures preserved	
9	Biotite-garnet schist and metagraywacke	BO-GA-MU-QZ-PL BO-MU-QZ-PL	PL is oligoclase to andesine	Morey, in prep
	Concretions	HO-GA-DO-ZO-MG-(CA)		
	Calcareous rocks	HO-BO-QZ-AB		
	Quartzose rocks	MU-BO-GA-ST-(CD-QZ)		
10	Iron-formation	Mn-Sp-Si-(Ch-Mg-He)		Schmidt, 1963
	Argillite	Mu-Qz-(Ch-Bo-Mg)		
	Mafic rocks	Cl-Ep-(Ab-Bo-Mu) Cl-Ep-Cz-Ab or Pl-Ca-Sh		
11	Argillite	Bo-Qz-Pl-(Cl)	Cl→Bo	Schmidt, 1963
12	Quartzose schist	BO-MU-QZ-(CD)		Morey, in prep
13	Quartzose schist	QZ-BO-GA-(MU-CD)		Morey, in prep
14	Quartzose schist	BO-MU-GA-ST-(CD-QZ)		Morey, in prep
15	Biotite schist	QZ-PL-BO-CO-GA-(MC-PX)	tonalite neosome; age of paleosome unknown Mu-Ch-Ep and Bo-Ga occur as retrograde minerals in cataclastic zones	Keighin et. al., 1972
16	Argillaceous rocks	Qz-Pl-Ch-Mu	"Republic node" Chlorite zone clastic feldspar	James, 1955
	Basic igneous rocks	Cl-Ep-Ac-Ab		
	Iron-formation	Gr-Ch-Si-He-Mg-(Mn-Sp)		
	Calcareous rocks	Ca		

Table 4 (cont'd.)

Locality	Rock Type	Minerals and/or Mineral Assemblage	Remarks	Reference
17	Argillaceous rocks	Qz-Pl-Bo-(Mu-Cl)	"Republic node" Biotite zone feldspar in part clastic	James, 1955
	Basic igneous rocks	Bo-Ep-Ab-(Cl-Ac-Ho)		
	Iron-formation	Qz-He-Mg-Sp-(Ch-Mn-Gu)		
	Calcareous rocks	Ca-(Tr)		
18	Argillaceous rocks	Qz-Pl-Mu-Bo-Ga	"Republic node" Garnet zone Pl is oligoclase	James, 1955
	Basic igneous rocks	Ho-Ab-(Bo-Ep)		
	Iron-formation	Qz-He-Mg-Gu-Ho-Ga		
	Calcareous rocks	Ca-Tr		
19	Argillaceous rocks	Qz-Ab-Mu-Bo-Ga-St	"Republic node" Staurolite zone	James, 1955
	Basic igneous rocks	Bo-Ho-Pl	Pl is andesine	
	Iron-formation	He-Mg-Gu-Ho-Ga		
	Dolomite	Ca-Tr-Di		
20	Argillaceous rocks	Qz-Pl-Mu-Bo-Si-(St)	"Republic node" Sillmanite zone	James, 1955
	Basic igneous rocks	Bo-Ho-Pl	Pl is andesine	
	Iron-formation	He-Mg-Px-(Gu-Ho)		
	Dolomite	Ca-Tr-Di		
21	Mafic volcanic rocks	Ab-Ep-Cx-Ac-Pl	relict textures preserved Cx→Ac Pl→Ep	Mursky and Hall, 1973
22	Trachyandesite	Pl-Ho-Bo-Qz	Pl occurs as relict phenoc	Medaris, et. al., 1973
23	Rhyolite and metasedimentary rocks	QZ-BO-MU-MC-AN		Lahr, 1972
	Basic volcanic flows	PL-HO-XP PL-HO-CU	CP occurs as relict grains	
	Metasedimentary rocks	QZ-BO-HO-PL-(EP) QZ-MC-BO-MU-(PL)		
	Basic volcanic rocks	QZ-PL-MU-BO-AN-(SI)		
	Metasedimentary rocks	GA-ID-SC-CX-HO	contact metamorphic aureole	
24	Mafic volcanic rocks	"amphibolite" locally well banded near granite contacts. Mafic volcanic greenstones having relict primary textures away from granite contacts.		LaBerge, 1973
25	"Quartz-eye" schist	Pl-Mu-Cl-(Co)	extensive metasomatic alteration	May, 1976
	Actinolite schist	Qz-Pl-Ac-Cl-Ga		
	Chlorite-garnet schist	Qz-Bo-Cl-(Ga)		
	Andalusite-biotite schist	Bo-An-Cl		
	Quartz-sericite schist	Qz-Mu-(Cl-Bo-An)		
26		Misc. drill holes report hornblende schist, sericite schist, and garnetiferous schist.		Dutton and Bradley, 1970
27	Mafic volcanic rocks	Ac-Qz-Pl-(Bo-Ca)	Watersmeet node of James, 1955 (Garnet grade)	Prinz, et. al., 1975
28	Mafic flows and associated volcanogenic rocks	low metamorphic grade (greenschist facies)	Peavy Node of James, 1955	Prinz, et. al., 1975
29	Mafic flows and mafic sills	low metamorphic grade (greenschist facies)	Px→Tr or Ac	Prinz, et. al., 1975

See Appendix I for explanation of symbols and Figure 6 for localities. Upper case symbols refer to mineral assemblages; lower case symbols to diagnostic minerals.

that Terrane III deposition in areas well to the south of the Terrane I-Terrane II boundary was characterized by volcanism and little attendant clastic sedimentation.

Deposition of Terrane III rocks was terminated or closely followed by an orogenic event – Penokean Orogeny of Goldich et al. (1961) – that involved deformation and metamorphism of the stratified rocks. The resulting structural and metamorphic features are closely related to the nature of the basement rocks. Where Terrane III rocks overlie granitic rocks of Terrane II, they are only slightly if at all metamorphosed, and the surface between the basement and overlying rocks appears to be relatively undisturbed (Sims, 1976b; Morey, 1977). However where Terrane III rocks overlie metasedimentary and metavolcanic rocks of Terrane II or gneissic rocks of Terrane I, they are extensively metamorphosed and both the supracrustal rocks and the underlying rocks are complexly infolded. Deformation was manifested principally by vertical tectonic processes leading to the development of fault-bounded blocks (Cannon, 1973) or mantled gneiss domes (Morey and Sims, 1976; Sims, 1976b).

Most igneous rocks that intrude Terrane III stratified rocks postdate the Penokean Orogeny. They have been well dated in Michigan and Wisconsin at about 1900 Ma (Aldrich et al., 1965), 1800 Ma and 1500 Ma (Van Schmus, 1976). Various volcanic units within the terrane have been dated at 1950 Ma (Banks and Van Schmus, 1972), 1900 Ma (Banks and Rebello, 1969) and at about 1870 Ma (Sims, 1976c). In northern Minnesota, Terrane III rocks overlie mafic dykes dated at about 2000 Ma (Hanson and Malhotra, 1971). Thus it appears that most of this terrane was deposited about 1900 to 2000 Ma ago. However in east-central Minnesota, definite Terrane III rocks are underlain by a complexly folded, migmatitic unit radiometrically dated at about 2000 to 2100 Ma (Goldich, 1973). The stratigraphic position of the metamorphosed sedimentary rocks in this migmatite complex is unknown. They may represent remnants of Aphebian strata equivalent to or older than typical Terrane III rocks, or they may be Terrane I or Terrane II metasedimentary rocks caught up in a younger orogenic event. However, inasmuch as their structural attributes show neither Terrane I nor Terrane II affinities (Morey, in prep.) the metamorphosed rocks are considered here to be part of Terrane III.

Data currently available on the metamorphic attributes of Terrane III rocks are summarized in Table 4 and Figure 6. Much of the terrane has been only slightly, if at all, metamorphosed as indicated by well preserved sedimentary textures and minerals. For example, iron formation in the northern part of the area (Table 4, locality 1) contains chalcedony and greenalite (Floran and Papike, 1975), whereas associated argillaceous rocks (locality 2) contain IM_d and IM muscovite polytypes (Morey, 1969). Correlative iron-formation somewhat to the southwest (Table 4, locality 3) also contains these minerals as well as abundant minnesotaite and stilpnomelane (French, 1968). On the basis of the quartz-magnetite geothermometer, it is probable that quartz and magnetite pairs in these rocks were subjected to extensive isotopic homogenization at no more than 150°C (Perry et al., 1973), a temperature characteristic of either high grade diagenesis or low grade metamorphism. Associated argillaceous rocks (Table 4, locality 4) contain chlorite and 2M muscovite, minerals which persist for some distance to the south. Similar chlorite- and muscovite-bearing rocks also are exposed at localities 5 and 6 (Fig. 6) where the rocks are characterized by a slaty cleavage presumably formed in response to Penokean tectonism.

Although the regional metamorphic grade is generally low, Terrane III is characterized by a number of metamorphic nodes or areas of high metamorphic intensity that are defined by the progressive appearance of biotite, garnet, staurolite, and sillimanite. These nodes are one of the most distinctive features of Terrane III metamorphism. Their areal extent in Michigan and Wisconsin and their general mineralogic characteristics were first outlined by James (1955), and subsequent studies (Cannon, 1973; Cannon and Klasner, 1972) have shown that their distribution is not related to the regional structure or to exposed granitic intrusions of Penokean age. Furthermore, the metamorphic highs do not appear to be related to depth of burial inasmuch as the rocks have been neither tilted nor deformed since they were metamorphosed. Instead, the various isograds are nearly isotherms and the present peneplaned erosion surface is nearly an isobaric surface (Cannon, 1973).

Sims (1976b) recognized that the various metamorphic highs are associated spatially with uplifted blocks or domes of Terrane I gneissic material. Therefore he suggested that Penokean tectonism and metamorphism were related to vertical tectonic processes involving the reactivation of underlying Terrane I rocks. Reactivation led to the formation of a number of fault-bounded blocks and/or gneiss domes characterized by cataclasis, recrystallization, and in places metamorphic segregation and/or partial melting. Bedding in Terrane III strata that surround and overlie the uplifted blocks is virtually parallel to cataclastic foliations within the uplifted blocks, implying that folding and uplift occurred nearly contemporaneously. The occurrence of higher grade metamorphic rocks along the flanks of uplifted Terrane I blocks suggests that deformation was accompanied by relatively high heat flow, which produced the metamorphic patterns now observed.

Local differences in the size and shape of the various metamorphic nodes shown in Figure 3 most likely are related to the relative plasticity of the basement rocks during deformation. In the development of gneiss domes, the basement rocks behaved plastically, and textural reconstitution was accompanied by dynamothermal metamorphism and local partial melting. In these situations the associated metamorphic isograds are widely separated and define nodes which cover moderately large areas (Table 4, localities 7-11 and 17-20). In contrast, uplift of fault-bounded blocks took place under less intense metamorphic conditions and was characterized by brittle deformation with only minor recrystallization and anatexis. In these situations, the associated metamorphic isograds are compressed and define elongate nodes which in places parallel major faults (Table 4, locality 27).

Very little is known regarding the metamorphic attributes of Terrane III volcanic rocks in north-central Wisconsin. However in places (Table 4, locality 25) these rocks have been metamorphosed to the lower amphibolite facies (Dutton and Bradley, 1970) and in other places (Table 4, locality 26) to at least the greenschist facies. Consequently the entire unit is assigned on Figure 3 to the greenschist-amphibolite facies, undivided.

Although most of the mineralogical data summarized in Table 4 relates to the Penokean Orogeny, some evidence suggests that parts of the terrane were metamorphosed to varying degrees by both pre- and post-Penokean events. For example, typical pre-Terrane III rocks in parts of east-central Minnesota (Table 4, locality 15) are characterized by prograde mineral assemblages indicative of the upper amphibolite facies. These rocks in turn have undergone cataclasis, and the cataclastic zones contain a variety of retrograde minerals, including biotite and garnet, that formed during the Penokean Orogeny (Morey, in prep.). Thus the cordierite-bearing mineral assemblages appear to record a discrete metamorphic event which preceded the Penokean Orogeny and perhaps deposition of Terrane III strata. The fact that typical Terrane III strata in parts of northern Michigan unconformably overlie basement rocks that were deformed, metamorphosed, and intruded by igneous rocks at

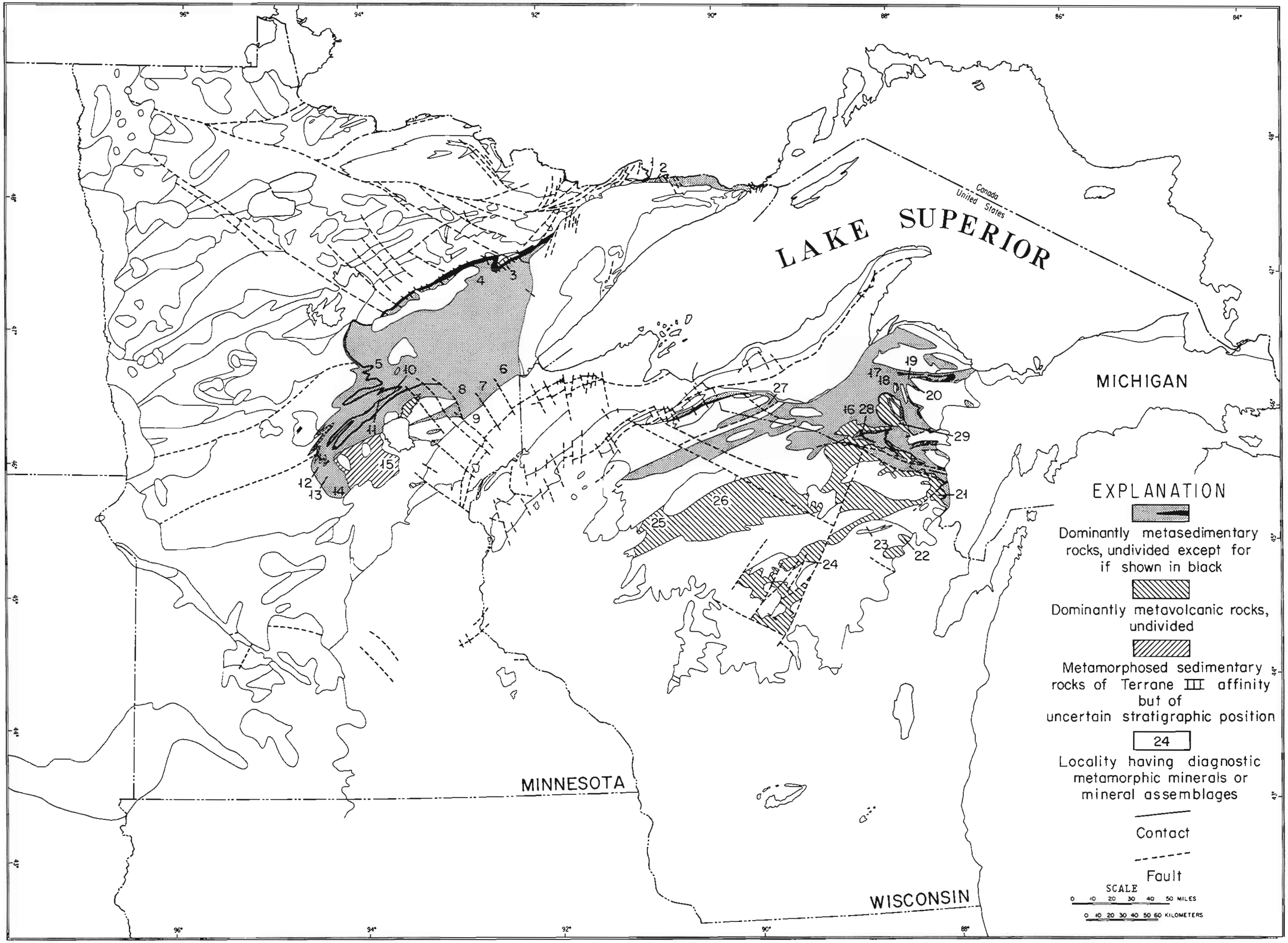

Figure 6. Bedrock geologic map of the Lake Superior region showing the known or inferred distribution of rocks assigned to Terrane III — Intracratonal supracrustal rocks. See Table 4 for inventory of diagnostic metamorphic minerals and mineral assemblages.

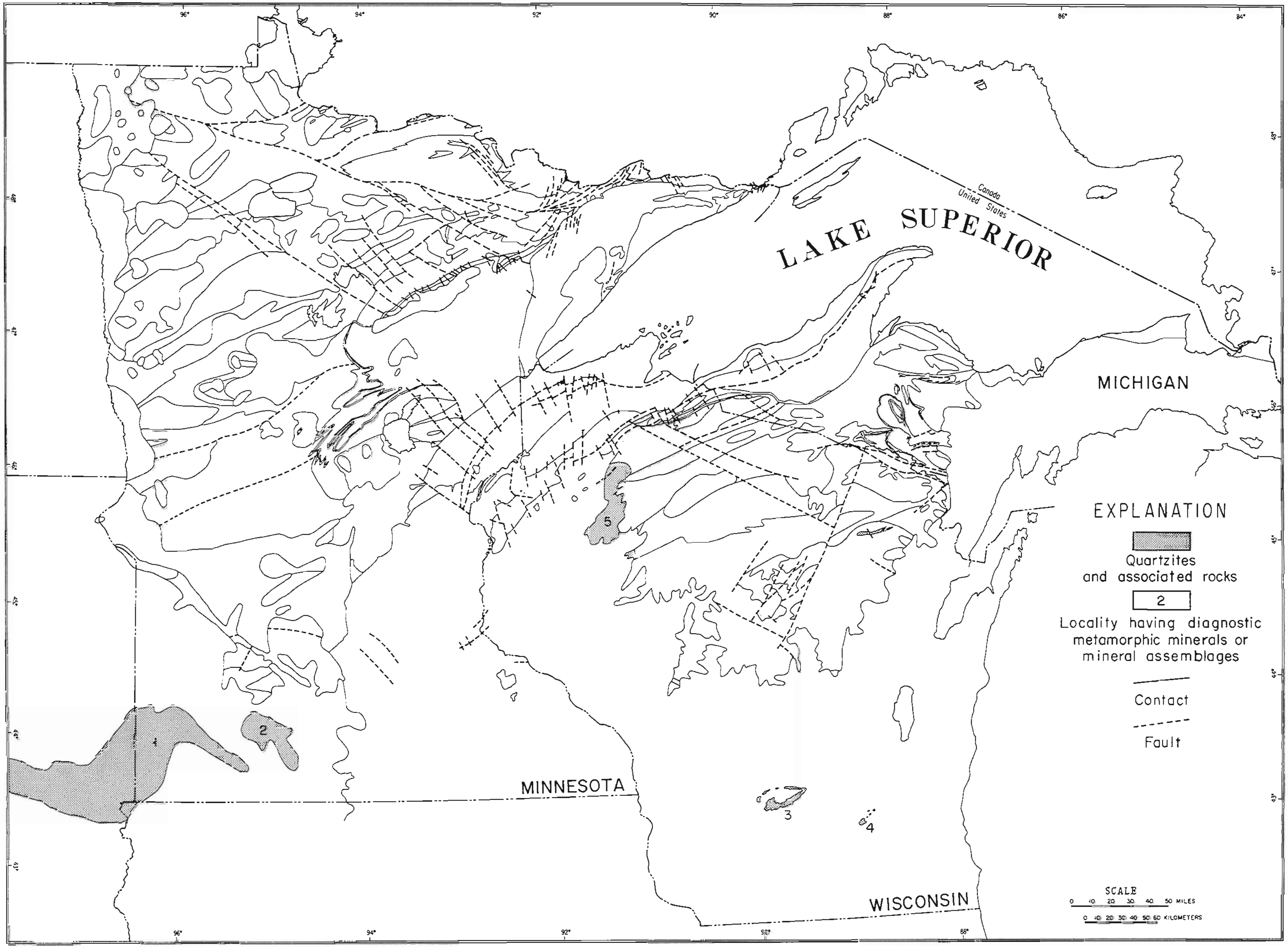

Figure 7. Bedrock geologic map of the Lake Superior region showing the distribution of rocks assigned to Terrane IV — Platform-type supracrustal rocks. See Table 5 for inventory of diagnostic metamorphic minerals and mineral assemblages.

about the same time (2100 Ma) is consistent with the concept of a widespread period of metamorphism which occurred prior to deposition of Terrane III strata in the Lake Superior region (Van Schmus, 1976).

Rocks of Terrane III also were affected by several post-Penokean metamorphic events. Van Schmus and Woolsey (1975) have shown by detailed radiometric studies that in northern Michigan these rocks were subjected to a low grade metamorphic event about 1700 to 1650 Ma ago. Textural and mineralogical evidence which may be associated with this event is generally lacking, but in east-central Minnesota (Table 4, localities 7 to 9) elongate muscovite grains yielding K-Ar ages of about 1610 to 1630 Ma (Goldich et al., 1961) define a secondary fabric which is superposed on the metamorphic fabric formed during the Penokean Orogeny (Morey, in prep.). Similarly many of the post-Penokean igneous bodies emplaced in Terrane III rocks at around 1800 Ma and 1500 Ma in northeastern Wisconsin have narrow contact metamorphic aureoles characterized by assemblages indicative of the pyroxene-hornfels facies (Table 4, locality 23).

Terrane IV — Platform-Type Supracrustal Rocks

Locally deformed and metamorphosed quartzite and lesser amounts of conglomerate and Al-rich mudstone comprise most of Terrane IV. However in some places, such as south-central Wisconsin, the quartzites are underlain by ignimbrites and tuff-breccias of rhyolitic composition and are overlain by various kinds of iron-rich and argillaceous strata. Although precise stratigraphic relationships between these rocks are uncertain, the underlying and overlying rocks also are included in Terrane IV. Insofar as known, this terrane, which is at least several thousand metres thick, lies entirely on gneissic rocks of Terrane I (Fig. 7).

Not all of the rocks assigned to Terrane IV have been deformed or metamorphosed. For example, inferred Terrane IV rocks in northwestern Wisconsin (Table 3, locality 5) are generally flat-lying and contain mudstone units rich in diaspore, illite, kaolinite and quartz; these rocks are classified as "unmetamorphosed" on Figure 3. In contrast, presumably equivalent rocks in south-central Wisconsin (Table 5, localities 3 and 4) have been folded about east-northeast trending axes (Dalziel and Dott, 1970) and contain minerals — quartz, pyrophyllite, muscovite, and hematite — indicative of greenschist or higher facies. Metamorphism may have been more intensive as indicated by the tentative identification of andalusite (Weidman, 1904). However, inasmuch as andalusite has not been verified by modern petrologic methods (Dalziel and Dott, 1970), these rocks are assigned to the greenschist facies on Figure 3. Similarly, Terrane IV rocks in southwestern Minnesota, have been folded about north-northeast-trending axes (Austin, 1972), and the probable presence of pyrophyllite (Berg, 1938) in Al-rich mudstone units (Table 4, localities 1 and 2) suggests that they also have been metamorphosed to the greenschist facies. However, Al-rich rocks in other parts of the terrane in southwestern Minnesota contain only kaolinite and illite and these were therefore assigned to the sub-greenschist facies (Fig. 3).

Although the transformation from an Al-rich precursor to pyrophyllite is the most diagnostic indicator of metamorphism in Terrane IV rocks, it appears that metamorphism did not everywhere take place at the same time. Van Schmus (1976) suggested that the Terrane IV rocks in south-central Wisconsin were metamorphosed at about 1650 Ma (see also Dott and Dalziel, 1972), whereas similar rocks in southwestern Minnesota yield K-Ar mineral ages of approximately 1200 Ma (Goldich et al., 1961).

Terrane V — Rocks of the Midcontinent Rift System

The final event in the formation of the Precambrian crust in the Lake Superior region was the development of the so-called "Midcontinent rift system" (King and Zietz, 1971). Rifting was accompanied by the massive upwelling of mantle-derived magmas with the solidification of mafic plutonic rocks at depth and widespread volcanism and clastic sedimentation at the surface. Rocks associated with this period of rifting comprise Terrane V.

Terrane V is particularly well developed in northeastern Minnesota where a variety of mafic plutonic rocks and their volcanic roof rocks are exposed, for 250 km along the north shore of Lake Superior (Fig. 8). These rocks disappear beneath the Paleozoic strata of the Midcontinent region 160 km south of the west end of Lake Superior, but they continue southward in the subsurface for more than 1000 km as a belt of basaltic volcanic rocks 40 to 85 km wide (King and Zietz, 1971). These rocks yield the so-called "Midcontinent Gravity High" characterized by one of the largest positive gravity anomalies in the United States. Terrane V rocks also continue east-northeastward from the west end of Lake Superior in a broad belt extending across northern Wisconsin and Michigan (White, 1966) before turning southward and disappearing again beneath Paleozoic strata of the Michigan basin (Oray et al., 1973).

Rocks within the rift system can be arbitrarily divided into three lithotectonic assemblages which partially overlap in space and time. The oldest assemblage consists predominantly of sedimentary rocks and possibly coeval low-alumina, tholeiitic sills, which did not significantly deform or metamorphose their country rocks. The next assemblage is predominantly igneous and consists of at least two separate successions of lava flows and associated plutonic rocks that were emplaced over a relatively short span of time, about 1100 to 1000 Ma ago (Silver and Green, 1963; Chaudhuri, 1972). Pyroxene-phyric lava flows, pyroxene diabase sills and minor dykes, and gravity-segregated plutonic units — peridotite, anorthositic gabbro, and granophyre — comprise

Table 5

Inventory of diagnostic metamorphic minerals and mineral assemblages in Terrane IV — Platform-type supracrustal rocks

Locality	Rock Type	Minerals and/or Mineral Assemblages	Remarks	Reference
1	Al-rich argillite	Py-Qz-(Mu-Ka-It)	rock type also referred to as "catlinite" or "pipestone"	Berg, 1938
2	Mudstone	Mu-He-Da-Qz		Miller, 1961
3	Argillaceous material	Py-Qz-(Mu-He)	pronounced recrystallization to "phyllite"	Dalziel and Dott, 1970
4	Slate or quartzite	QZ-CL-MU	"phyllite" in quartzite	Dott and Dalziel, 1972
	Argillite	Mu-rich		
5	Al-rich argillite	Da-Qz-(Ka-It)	Barron "pipestone"	Dott and Dalziel, 1972

See Appendix I for explanation of symbols and Figure 7 for localities. Upper case symbols refer to mineral assemblages; lower case symbols to diagnostic minerals.

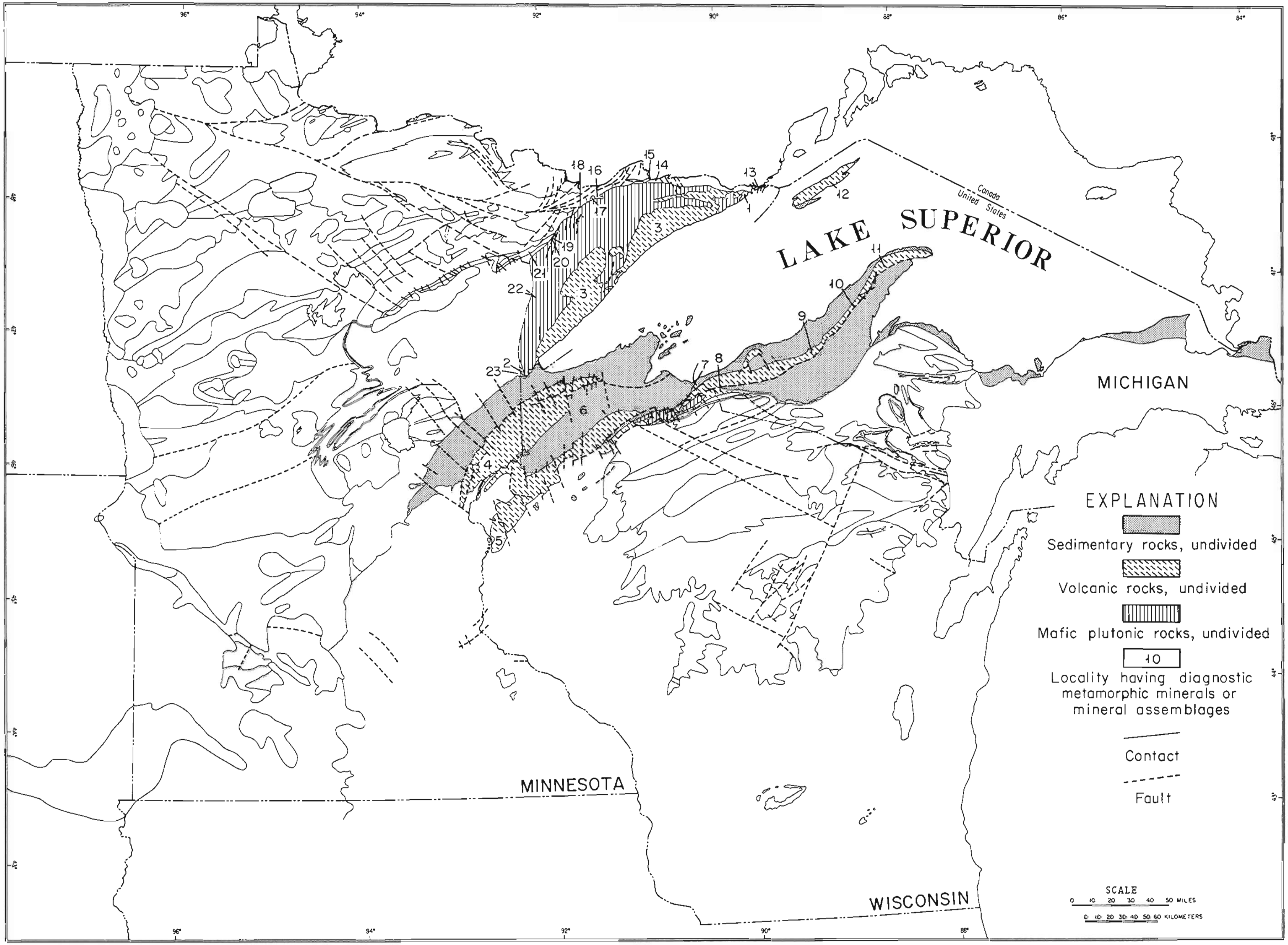

Figure 8. Bedrock geologic map of the Lake Superior region showing the distribution of rocks assigned to Terrane V – Midcontinent rift system. See Table 6 for inventory of diagnostic metamorphic minerals and mineral assemblages.

the older suite. The younger suite consists of plagioclase-phyric lava flows, olivine diabase dykes and minor sills, and troctolitic-gabbroic plutonic rocks. Emplacement of these contrasting igneous rock suites appears to have been related to the progressive development of surface and subsurface void spaces produced by extensive faulting, which increased in frequency and intensity with time (Weiblen and Morey, 1976). The uppermost litho-tectonic assemblage consists of two suites of clastic sedimentary rocks of alluvial to fluvial origin. The older suite, which locally is intercalated with lava flows, consists of lithic sandstone and shale that were deposited in a number of faulted-bounded basins along the axis of the rift (Morey, 1974). The younger suite consists of arkosic and quartzose sandstone deposited in large, half-graben-like basins along the flanks of the rift (Morey, 1972). These predominantly sedimentary assemblages mark the gradual cessation of crustal separation and magmatism. However, dominantly vertical faulting continued intermittently throughout the time of active sedimentation and into the Paleozoic Era.

Available data on the metamorphic attributes of Terrane V rocks are summarized in Table 6 and Figure 8. Many of the stratified rocks and particularly the lava flows have undergone low-rank burial metamorphism ranging from the zeolite facies through the prehnite-pumpellyite facies to, locally, at least, the lower part of the greenschist facies (Table 6 and Fig. 8). In general, lava flows metamorphosed to the greenschist facies are more or less pervasively altered, whereas those metamorphosed to the subgreenschist facies are altered only along vesicular flow tops and fracture zones. The altered flow tops and fracture zones are characterized by an almost complete transformation to monomineralic metadomains (Jolly and Smith, 1972) consisting of one of the following minerals: albite, quartz, epidote, or various zeolites or carbonates. Although the lava flows have been completely transformed in these metadomains, there is a gradation away from them through incompletely reconstituted rocks containing albitized feldspar and oxidized mafic minerals to rocks displaying few signs of mineralogical adjustment. Thus metamorphism appears to have resulted from fluids that migrated along permeable channelways following extrusion of a considerable thickness of flows (Stoiber and Davidson, 1959). Several detailed chemical studies have shown that the lava flows were the source of the chemical constituents in the altered zones with the possible exception of K_2O and water (Jolly and Smith, 1972; Scofield, 1976).

In general, it is possible to divide the lava sequences exposed along both the north and south shores of Lake Superior into a number of metamorphic zones characterized by thermal intensity that increases with depth. For example, the lava flows at locality 11 (Table 6) in northern Michigan have been divided into three metamorphic zones having broadly gradational boundaries (Jolly and Smith, 1972). The first or uppermost zone is characterized by chlorite, albite, prehnite, laumontite, analcime, and sphene and is considered to be within the zeolite facies as defined by Coombs et al. (1959). A second zone occupying the middle half of the section is characterized by pumpellyite, albite, chlorite, prehnite, and quartz and is considered by Jolly and Smith to be within the prehnite-pumpellyite facies of Coombs (1960). The lowermost flows at this locality constitute a third zone containing albite, epidote, quartz, and lesser pumpellyite — minerals also considered by Jolly and Smith to be indicative of the prehnite-pumpellyite facies. Although these zones strictly apply only to locality 11 of Figure 8 the earlier more regional studies of Stoiber and Davidson (1959) suggest that similar zones are prevalent throughout the entire stratigraphic succession of exposed lava flows in northern Michigan.

Mineral assemblages similar to those in Jolly and Smith's (1972) uppermost zone in northern Michigan also occur throughout much of the lava sequence exposed in Minnesota along the north shore of Lake Superior (Green, 1972). However the presence of zeolites such as stilbite, heulandite, scolecite, and mordenite suggests that the upper parts of this sequence were less intensely metamorphosed than the uppermost lava flows exposed in Michigan and Wisconsin. Similarly, Green (1972) suggested that the basal pyroxene-phenocrystic lavas in northern Minnesota (Table 6, localities 1 and 2) were metamorphosed to the greenschist facies as part of the same period of burial metamorphism. If so, burial metamorphism along the north shore occurred at stratigraphic levels both deeper and higher than that which occurred along the south shore of Lake Superior. However Hubbard (1975b) suggested that the basal pyroxene-phenocrystic lavas in Wisconsin and adjoining parts of Michigan (Table 6, locality 8) were metamorphosed to the greenschist facies before the overlying and less metamorphosed lavas were deposited. Thus it is not clear whether Terrane V rocks have been subjected to one or two distinct periods of metamorphism.

Lithic sandstones intercalated with and overlying the lava flows (Table 6, locality 6) also have been metamorphosed as indicated by extensive recrystallization, particularly along framework grain-matrix boundaries, by the oxidation of lithic rock fragments, and by the progressive transformation of mixed-layer/illite-montmorillonite to chlorite and well-ordered sericite (Morey, 1974). Thus it appears that these rocks were subjected to the same, although less extreme, metamorphic processes that affected the bulk of the underlying lava flows. In contrast, the arkosic and quartzose sedimentary rocks deposited in half-graben-like basins along the flanks of the rift system contain no textural and/or mineralogic evidence indicative of burial metamorphism (Morey, 1972). Because metamorphism of the lava flows and associated lithic sandstone must have taken place beneath an extensive thickness of overlying strata, it must be concluded that a major period of uplift and erosion occurred before the arkosic and quartzose sandstone was deposited.

Although there is no evidence that the plutonic rocks comprising a major part of the Midcontinent rift system have been metamorphosed, they have markedly metamorphosed their country rocks. Contact metamorphic aureoles several tens of metres to several hundreds of metres wide are particularly well developed around intrusive bodies, most of which are too small to be shown on Figures 2 and 3. Although variable in width, all metamorphic aureoles are characterized by granulitic, granoblastic, or hornfelsic textures and metamorphic assemblages assignable to the pyroxene-hornfels or hornblende-hornfels facies of contact metamorphism.

The contact metamorphic effects of Terrane V plutonic rocks on Terrane III argillaceous sedimentary rocks and iron formations have been extensively studied in northeastern Minnesota (Gundersen and Schwartz, 1962; French, 1968; Morey et al., 1972; Bonnichsen, 1975; Floran and Papike, in press). In general, metamorphism of the iron formation is pronounced and the aureoles (Table 6, localities 15, 20, and 21, and Fig. 8) can be divided into a number of mineralogic zones characterized by: (1) the appearance of minnesotaite and stilpnomelane, (2) the disappearance of the layered silicates and carbonates and the appearance of minerals of the grunerite-cummingtonite series, (3) the appearance of hedenbergite and fayalite, and (4) the disappearance of most of the prograde amphiboles and the appearance of orthopyroxene and augite. Zone 4 is also characterized by extensive retrograde reactions involving exsolution, inversion, oxidation, and hydration. Early-formed pigeonite exsolved augite, and was inverted to orthopyroxene which also exsolved augite. Early-formed augite exsolved

Table 6

Inventory of diagnostic metamorphic minerals and mineral assemblages in Terrane V — Midcontinent rift system

Locality	Rock Type	Minerals and/or Mineral Assemblages	Remarks	Reference
		I. Regional Metamorphism		
1	Pyroxene-phenocrystic basalt		primary textures and minerals preserved. Px→Ac Pl saussuritized	Green, 1972
	Amygdule minerals	Qz-Ph-Ca-Ep-Ch		
2	Pyroxene-phenocrystic basalt	(same as locality 1)		Kilberg, 1972
3	Mafic to felsic lava flows		primary textures and minerals preserved Pl and Px unaltered opaque minerals & pigeonite partly oxidized fresh undevitrified volcanic glass present locally	Green, 1972
	Amygdule minerals	Kf-Lu Ca-Lu-Sb-Hu-Th-So-Mo-(Al-Nt-Ms-Ap) Qz-Cl native copper		
4	Mafic to felsic flows		primary textures and minerals preserved Px→Ac Pl saussuritized	Morey and Mudrey, 1972
	Amygdule minerals	Qz-Ca-Cl-Ep-Ph-Lu		
5	Amydgule minerals	Ep-Qz-Ph-Cl-Ca-Lu		Stoiber and Davidson, 1959
6	Sandstone and shale		recrystallized illite-montmorillionite chlorite	Morey, 1974
7	Mafic lavas		Px→Ho	Hubbard, 1975b
8	Pyroxene-phenocrystic basalt		primary textures and minerals preserved Pl saussuritized Pl→Ab, Cl & Ep or Pl→Cl & Ep Px→Ac & Cl ilmenite→leucoxene	Hubbard, 1975b
	Amygdule minerals	Qz-Ep-Cl-Ca Ep-Cl-Ph-Pu-Qz-Ab Cl-Ep-Qz-chalcedony-Ca-Lu-Pu-Ph-Zo-Cz		
9	Amygdule minerals in mafic lavas	Qz-Ph-Ca-Ep-Cl		Stoiber and Davidson, 1959
10	Mafic to felsic flows	"prehrite-pumpellyite facies"	Au→Cl Pl→Ab+Mu Px→Pu&/or Ep	Scofield, 1976
11	Mafic to felsic flows	Cl-Lu-Ph-Al-Ab-Sh Cl-Sh	Laumonite zone	Jolly and Smith, 1972
		Pu-Cl-Ph-Qz-Ab-Sh Pu-Sh-Qz	Pumpellyite zone	
		Ep-Pu-Ch-Qz-Sh Ep-Sh-(Qz)	Epidote zone	
12	Amygdule minerals	Ep-Qz-Ph-Cl-Ca-Lu		
		II. Contact metamorphism of country rocks		
13	Argillite and graywacke	QZ-PL-BO-(HY)	HY present only in proximity to contacts with igneous bodies	Mudrey, 1973
14	Quartzitic and quartzo-feldspathic rocks	BO-MU-QZ BO-MU-PL-QZ BO-MU-MC-PL-QZ	near contacts with igneous bodies	Morey, 1969
	Argillite and graywacke	AN-PL-QZ AN-PL-MU-QZ AN-CO-BO-QZ CO-BO-MU-PL-QZ CO-HY-PL-QZ HY-PL-QZ HO-PL-QZ HO-BO-PL-QZ		
	Carbonate concretions	CA-EP-CL-QZ DI-GA-PL-QZ DI-GA-CA-QZ		

Table 6 (cont'd.)

Locality	Rock Type	Minerals and/or Mineral Assemblages	Remarks	Reference
15	Iron-formation	Qz-Mn-Sp-Ca-Do-Ak-Sd-Mg-(Ch-Gr-He-Pi)	well away from contacts with igneous bodies	Floran and Papike, in press
		Qz-Gu &/or Cu-Ol-Hd-Ac	Grunerite zone	
		Hy-Qz-Hd-Ol-(Ac-Gu)	Ferro-hypersthene zone immediately adjacent to contact with igneous bodies Ac and Gu retrograde minerals	
		OP-OL-IL-PL-AU-(CU) OL-QZ-MG	immediately adjacent to contact with igneous body.	Simmons, et. al., 1974
16	Metasedimentary rocks	HO-QZ-(MC) PL-HO-(HY-MG or IL)		Green, 1970
17	Granitic rocks	Pl with antiperthitic patches of Mc Ferromagnesian minerals altered to: Ho-Bo-Mg, or Ho-Au-Hy-Bo-Mg	HY→BO+AC	Green, 1970
18	Metabasalt	PL-HO-AU-HY-MG-IL HO-DI-PL-BO-MG		Green, 1970
19	Argillite	QZ-MU-BO-PL-(CL-PI)	away from contacts with igneous bodies	Renner, 1969
		QZ-BO-PL-(CO-DI-PO) HY-PL-CO-(BO-IL)	near contacts with igneous bodies HY→AC	
	Carbonate concretions	CA-GA-WO		
20	Iron-formation	QZ-MG-HY-HD-OL-(CU&/or GU)	OL+QZ→HY OL→CU&/ or GU HY→CU&/or GU	Bonnichsen, 1975
	Quartzite	MU-CL-AB-QZ MU-CL-QZ MU-BO-AB-QZ MU-BO-CL-PL-QZ		Griffin and Morey, 1969
	Iron-formation having abundant quartz	QZ-CU QZ-HO CU-HO CU-HY CU-HY-QZ CU-HO-GA CA-HD-CU CU-AC CU-AC-HO	HY→CU	
	Iron-formation having minor quartz	CU-HO CU-HY-QZ CU-OL CU-HY-OL CU-OL-BO CU-HD HO-GA	OL→CU	
	Calcareous rocks	CA-DI CA-EP CA-GA CA-EP-CL CA-DI-GA		
	Argillaceous rocks	MU-CL-AB-QZ MU-CL-QZ MU-BO-AB-QZ MU-BO-CL-PL-QZ CO-BO-MU-PL-QZ		
21	Iron-formation	Qz-Gu &/or Cu-(Ca-Ac)	away from contacts with	French, 1968
		Qz-Gu &/or Cu-Ac-Hd-Ol	igneous bodies	
		Qz-Hd-Ol-Hy	Ac&Gu&/or Cu retrograde near contacts with igneous bodies	
22	Argillite	"cordierite-bearing"		Mainwaring 1975
23	Argillite	"cordierite-bearing"		Schwartz, 1942

See Appendix I for explanation of symbols and Figure 8 for localities. Upper case symbols refer to mineral assemblages; lower case symbols to diagnostic minerals.

pigeonite, whereas prograde cummingtonite exsolved actinolite. Subsequently olivine, orthopyroxene, and augite reacted with plagioclase and water to form various amphiboles (Simmons et al., 1974).

Contact metamorphic effects in Terrane III argillaceous rocks are much less pronounced than those in associated iron formations. Only in the immediate proximity of some of the larger igneous bodies do mineral assemblages contain quartz, biotite, plagioclase, pyrrhotite, and minor to locally abundant amounts of pyroxene, andalusite, and cordierite; in most locations, however, the latter two minerals have been altered to chlorite and sericite. Rocks near smaller igneous bodies and those several metres from large igneous bodies contain biotite, muscovite, quartz, plagioclase, and pyrite; rocks distant from any heat source contain chlorite rather than biotite.

The outer limits of contact metamorphism in both the iron formations and the argillaceous rocks are difficult to establish. In argillaceous rocks, the outer limit appears to correspond to the transformation of $1M_d$ and 1M muscovite to the 2M polytype (Morey, 1969). In the iron formations mineral assemblages attributable to contact metamorphism (Table 6, locality 15) are similar to those that formed by either high grade diagenetic or low grade metamorphic processes (Table 4, locality 3) during an earlier period of regional metamorphism.

The distribution of metamorphic mineral assemblages, particularly in the iron formations, indicates that prograde metamorphism was essentially isochemical except for the loss of carbon dioxide and water. Retrograde metamorphism was common, as evidenced partly by reactions between anhydrous silicates and partly by reactions involving the formation of hydrous minerals. Temperatures attained during prograde metamorphism have been estimated (Bonnichsen, 1975) to have been between 650° to 750°C at locality 20 (Table 6 and Figure 8). The occurrence of numerous irregular granitic veins derived by local partial melting in argillaceous rocks at nearby locality 19 (Table 6 and Fig. 8) suggests that the minimum temperature was between 675° and 700°C if the pressure was between 1.5 and 3 kb (Bonnichsen, 1972, 1975). These estimates are in general agreement with the estimate (600° to 675°C, 1.5 to 2.5 kb) made by Green (1970) for nearby metamorphosed granitic rocks, the estimate (800°C, 2 kb) made by Simmons et al. (1974) for contact metamorphosed iron formation at locality 15 and the estimate (650°C) for highly metamorphosed iron formation (Perry et al., 1973). Temperatures at the outer edges of the contact metamorphic aureoles are more difficult to establish, but oxygen-isotope studies by Perry et al. (1973) suggest that assemblages containing minnesotaite and stilpnomelane formed at temperatures of no more than 150°C.

SUMMARY

From the point of view of the terrane concept presented above, the fundamental geologic map of the Lake Superior region is relatively simple (Fig. 9). It consists of Terranes I, II, and V, each representing a primary block of the craton formed by the upwelling of vast quantities of mantle-derived magmatic materials. The gneissic rocks of Terrane I record a long span of geologic history extending back to one of the earliest identifiable events in earth history at about 3800 to 3700 Ma ago, which is a billion years before any major datable geological events in Terrane II rocks. Perhaps equally important is the fact that additional major plutonic events took place in the gneiss terrane almost a billion years later than in Terrane II. It also is now apparent that the gneiss terrane records a number of intermediate-age, metamorphic events not related to plutonism, which have been recognized only recently and mainly as the result of

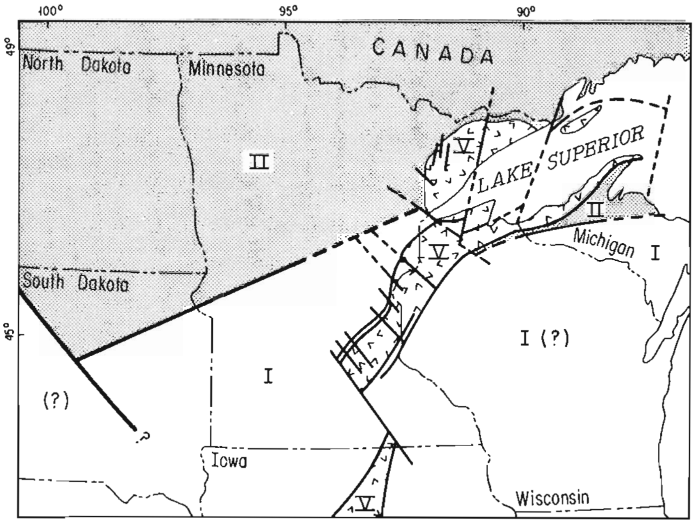

Figure 9. Geologic map showing Precambrian terranes that form the fundamental basement rocks of the Lake Superior region as they would appear if all basins and veneers of stratified rocks were stripped away. I, Gneiss Terrane; II, Greenstone-granite Terrane; and V, Midcontinent rift system.

chronometric investigations. Generally, petrologic data bearing on the metamorphic expression of these various events are lacking, a fact emphasized by the relatively sparse amount of definitive petrologic data regarding metamorphism of the gneiss terrane.

In contrast to Terrane I, Terrane II formed within a narrow range of a few hundred million years, commencing about 2750 to 2700 Ma ago, and has remained essentially stable since that time. This terrane is characterized by a number of linearly distributed supracrustal units which appear to have foundered or been down-folded between upwelling masses of granitoid rocks. Although metamorphism of the supracrustal rocks was related to emplacement of the granitoid rocks, their relatively low grade implies that the supracrustal rocks were never very deeply buried following extrusion and/or deposition. In contrast, high temperature mineral assemblages are common in the granite-migmatite massifs of northern Minnesota, suggesting that these rocks originated well beneath the surface. Except for several intractable facts, it would be tempting to interpret the massifs as remobilized basement on which the supracrustal rocks were initially laid down. All the radiometric ages obtained thus far from rocks within the massif cluster within the same time span as do those from the bounding supracrustal and plutonic rocks. Furthermore, isotopic and minor- and trace-element data suggest that all of the rocks in the greenstone-granite terrane were derived from the same primitive, mantle or subcrustal sources. Although the nature of the substratum upon which the supracrustal rocks were deposited is conjectural, Morey and Sims (1976) have suggested that they formed on an oceanic crust adjacent to a pre-existing sialic craton composed of Terrane I gneiss. They further suggest that the Terrane II rocks were welded to the Terrane I rocks — to form a large coherent continental crustal segment — by the addition of voluminous granitic material about 2700 to 2600 Ma ago.

The prolonged or repeated mobility of Terrane I rocks has important implications for the evolution of Terrane III and Terrane IV rocks. Although both of these terranes comprise appreciable quantities of igneous material, they are essentially supracrustal areas of subsidence and sedimentation formed on the more fundamental blocks of Terrane I and

Terrane II. Metamorphism of Terrane III rocks is clearly related to the remobilization of underlying gneissic rocks, and it is tempting to suggest that the Terrane IV rocks were metamorphosed by similar, but less intense tectonic processes.

Although much younger, Terrane V rocks formed in response to tensional processes leading to the development of an incipient rift system. Metamorphism of this terrane is generally of low grade and seems to be related to the rapid emplacement and burial of thick successions of lava flows and sedimentary rocks. Thus except for local contact metamorphic processes, the emplacement of this terrane had little effect on the older rocks.

The evolution of the Precambrian rocks in the Lake Superior region as interpreted here is contrary to the concept of continental accretion of the North American continent. Instead of younger rocks being added in successive crudely zoned patterns as suggested by Engel (1963), data summarized here suggest that the craton probably attained approximately its present dimensions by the close of Archean time. Younger sedimentary and volcanic rocks were deposited in intracratonic basins evolved upon the Archean basement rocks, or were deposited in a rift system characterized by tensional forces which fractured and separated the older rocks. Thus since the advent of Proterozoic time, the Lake Superior region has been characterized by vertical rather than lateral accretionary processes. Indeed, if we carry the story forward, the Paleozoic and younger rocks of the Midcontinent region represent simply a recurrence of such supracratonal events.

ACKNOWLEDGMENTS

I wish to stress that the metamorphic map and interpretations presented here are at best a first approximation. I realize that this report probably contains errors resulting from only a limited knowledge of many parts of the region and the consequent problems of interpreting the many contributions to our understanding of the Lake Superior region. To those people whose data or ideas I have inadvertently misinterpreted or used without proper credit I apologize. However I owe much more than I can state to extended discussions with many colleagues, particularly Matt Walton and P.K. Sims. Ms. Joyce Triviski suffered through the preparation of several versions of the manuscript and Richard B. Darling drafted the illustrations.

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Note: Appendix 1 on following page.

Appendix 1

Explanation of Symbols used in Tables 2 through 6

AB – Albite (An_{0-15})
AC – Actinolite
AK – Ankerite
AL – Analcite
AM – Amphibole
AN – Andalusite
AP – Anthophyllite
AU – Augite
BO – Biotite
CA – Calcite
CD – Chloritoid
CH – Chamosite
CL – Chlorite
CO – Cordierite
CP – Calcic plagioclase (An_{15-100})
CU – Cummingtonite
CX – Clinopyroxene
CZ – Clinozoisite
DA – Diaspore
DI – Diopside
DO – Dolomite
EP – Epidote
FS – Feldspar
GA – Garnet
GR – Greenalite
GU – Grunerite
HD – Hedenbergite
HE – Hematite
HO – Hornblende
HU – Heulandite
HY – Hypersthene
ID – Idocrase
IL – Ilmenite
IT – Illite
KA – Kaolinite
KF – Orthoclase
KY – Kyanite
LU – Laumontite
MC – Microcline
MG – Magnetite
MI – Mica
MN – Minnesotaite
MO – Mordenite
MS – Mesolite
MU – Muscovite
NT – Natrolite
OL – Olivine
OP – Orthopyroxene
PH – Prehnite
PI – Pyrite
PL – Plagioclase
PO – Pyrrhotite
PU – Pumpellyite
PX – Pyroxene
PY – Pyrophyllite
QZ – Quartz
RU – Rutile
SA – Sapphirine
SB – Stilbite
SC – Scapolite
SD – Siderite
SH – Sphene
SI – Sillimanite
SO – Scolecite
SP – Stilpnomelane
ST – Staurolite
TH – Thomsonite
TR – Tremolite
WO – Wollasonite
ZE – Zeolite
ZO – Zoisite

METAMORPHISM IN THE EASTERN AND SOUTHWESTERN PORTIONS OF THE GRENVILLE PROVINCE

James H. Bourne[1]

Bourne, J.H., Metamorphism in the eastern and southwestern portions of the Grenville Province; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 315-328, 1978.

Abstract

In the eastern Grenville Province granulite facies rocks are present in seven geographically separate regions. Sillimanite and orthopyroxene assemblages are relatively common. Upper greenschist facies rocks are found in the region underlain by the Wakeham Group where chlorite, muscovite, and albite are stable. Lower and middle amphibolite facies assemblages occur on the Atlantic coast of Labrador and along the Grenville Front tectonic zone. The remainder of the eastern Grenville Province is either in the upper amphibolite facies or in undivided amphibolite facies.

Southwestern Grenville Province rocks of granulite facies and of facies transitional between amphibolite and granulite are abundant. Several kornerupine occurrences have been recorded in the Kazabazua area of the Gatineau Valley north of Ottawa. Lower to middle amphibolite facies rocks occur in the Grenville Front transition zone. The remainder of the area is in either upper amphibolite facies or undivided amphibolite facies. Prehnite and/or pumpellyite have developed along joint surfaces.

Kyanite is found only within 70 km of the Grenville Front and in the staurolite-bearing schist of the Wakeham Group rocks. Andalusite-bearing rocks and cummingtonite-plagioclase assemblages have not been found in the two parts of the Grenville Province examined. Hornblende-garnet assemblages are very common.

It is suggested that some of the charnockitic gneiss found associated with the anorthosite complexes in central Quebec and elsewhere may be the deformed and recrystallized equivalents of the adamellite complexes of Labrador, in which, due to the dry nature of the adamellitic starting material, the primary orthopyroxene has been preserved although the primary texture has been destroyed.

Résumé

Dans l'est de la province de Grenville, on rencontre le faciès granulite dans sept régions géographiques distinctes. Les assemblages à sillimanite et orthopyroxène sont relativement communs. On rencontre les roches du faciès schistes verts supérieurs dans la région du groupe de Wakeham, où la chlorite, la muscovite et l'albite sont stables. Les assemblages du faciès amphibolite inférieure et du faciès amphibolite moyenne se rencontrent sur la côte atlantique du Labrador, et le long de la zone tectonique du front de Grenville. Le reste des terrains de l'est de la province de Grenville sont classés dans le faciès amphibolite supérieure, ou dans la partie non-divisée du faciès amphibolite.

Dans le sud-ouest de la province de Grenville, les roches métamorphisées dans le faciès granulite et dans la partie intermédiaire du faciès amphibolite et du faciès granulite, sont abondantes. Dans la région de Kazabazua, située dans la vallée de Gatineau au nord d'Ottawa, on a signalé plusieurs fois la présence de kornérupine. Les roches métamorphisées dans le faciès amphibolite inférieure et le faciès amphibolite moyenne apparaissent dans la zone de transition du front de Grenville. Le reste de la région appartient soit à au faciès amphibolite supérieure, soit à la partie non-divisée du faciès amphibolite. De la prehnite ou de la pumpellyite, ou ces deux minéraux à la fois, se sont formées le long des surfaces des diaclases.

On ne trouve de cyanite que dans un rayon de 70 km à partir du front de Grenville, et dans les schistes à staurolite des roches du groupe de Wakeham. On n'a pas trouvé de roches à andalousite et d'assemblages à plagioclase et cummingtonite dans les deux secteurs étudiés de la province de Grenville. Les assemblages à grenat et hornblende sont très fréquents.

On suggère que certains des gneiss charnockitiques associés aux complexes anorthositiques du centre du Québec et d'ailleurs, sont peut-être les équivalents déformés et recristallisés des complexes adamellitiques du Labrador où, en raison de la très faible hydratation du matériau adamellitique de départ, l'orthopyroxène primaire s'est conservé, bien que sa structure primaire ait été détruite.

INTRODUCTION

Two geographically separate portions of the Grenville Province are considered in this report (Fig. 1). A reliability map for each region is included (Figs. 2 and 3). The areas of the Grenville Province compiled by other workers are also shown.

EASTERN GRENVILLE PROVINCE

The eastern Grenville Province comprises approximately 42 per cent (280 000 km^2) of the Grenville Province in Canada. Of this area, 41 per cent is in Quebec and 59 per cent is in Labrador and the Great Northern Peninsula of Newfoundland.

[1] Geological Survey of Canada

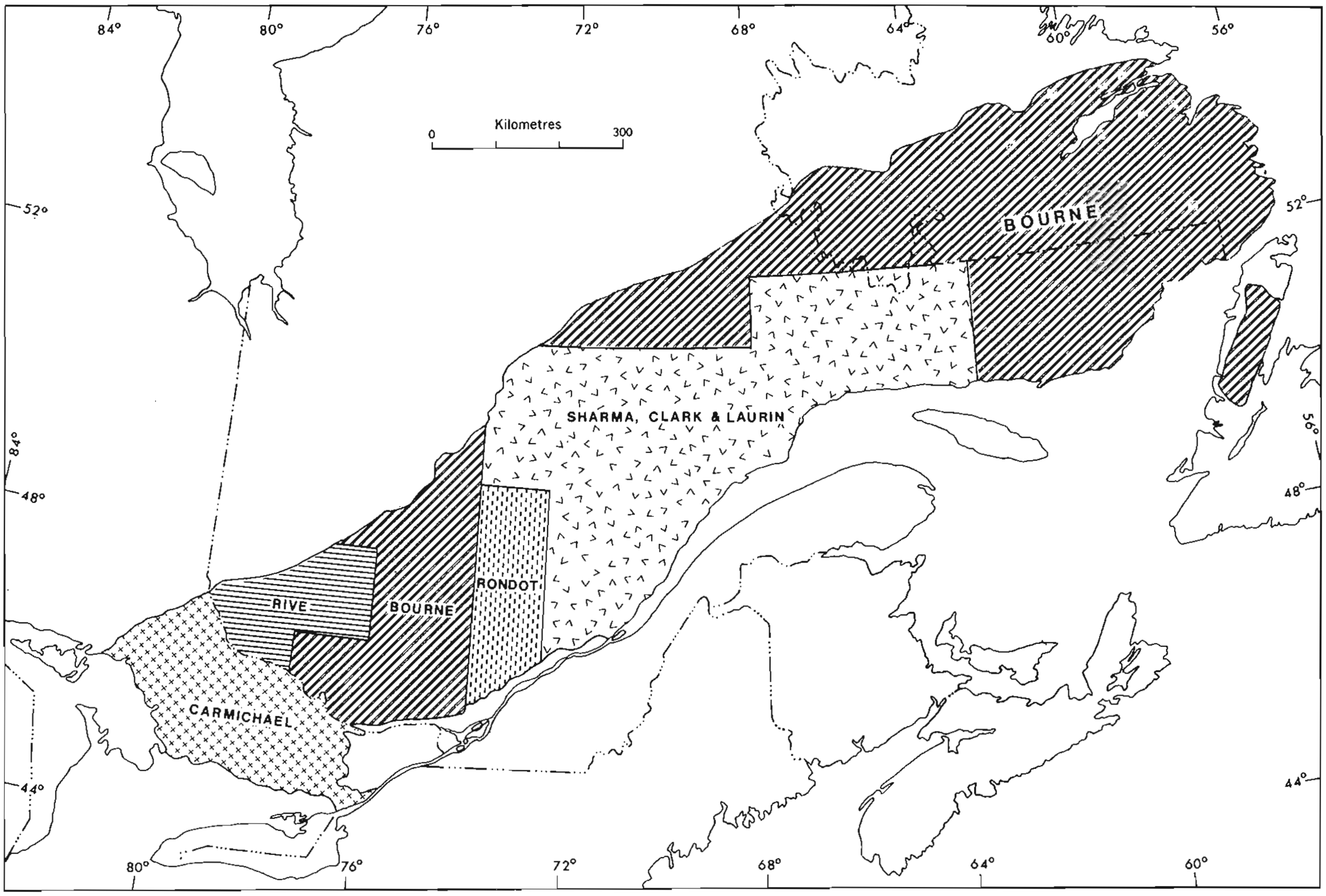

Figure 1. Outline of the Grenville Province showing the location of the eastern and southwestern regions of the Grenville Province described in this report as well as the parts of the Grenville Province compiled by others.

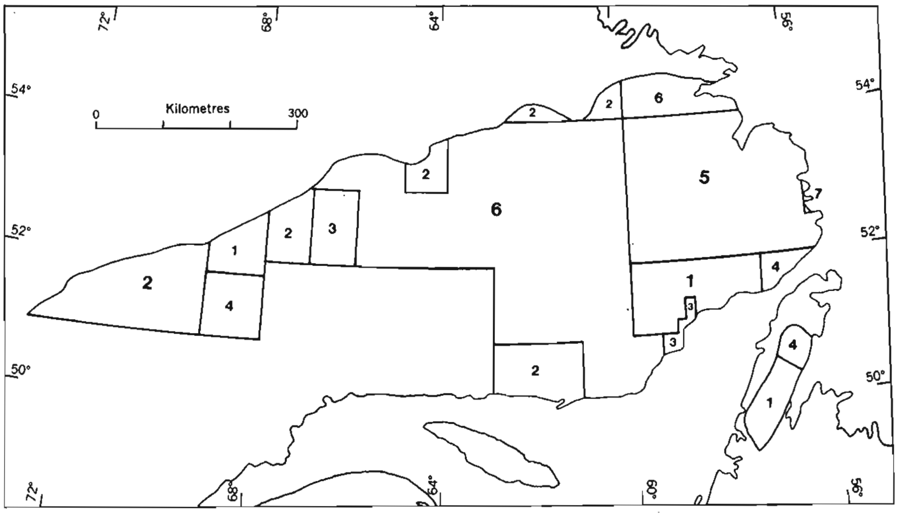

Figure 2. Sketch map of the eastern Grenville Province showing the reliability of the metamorphic data for different portions of the region.

1 — No information.

2 — No thin sections examined. Remarks on the mineralogy of different units made in the geological reports of the area, but reports do not contain any section pertaining specifically to metamorphism.

3 — As in (2) but reports include chapter dealing with metamorphism.

4 — Areas that have been subjected to a detailed petrologic investigation but not by the writer.

5 — Thin sections from this area examined by writer. Very sparse coverage and low sample density. Low reliability.

6 — As in (5) but sample density greater. Samples collected during 1:250 000 helicopter reconnaissance program.

7 — Detailed thin section collection examined by writer.

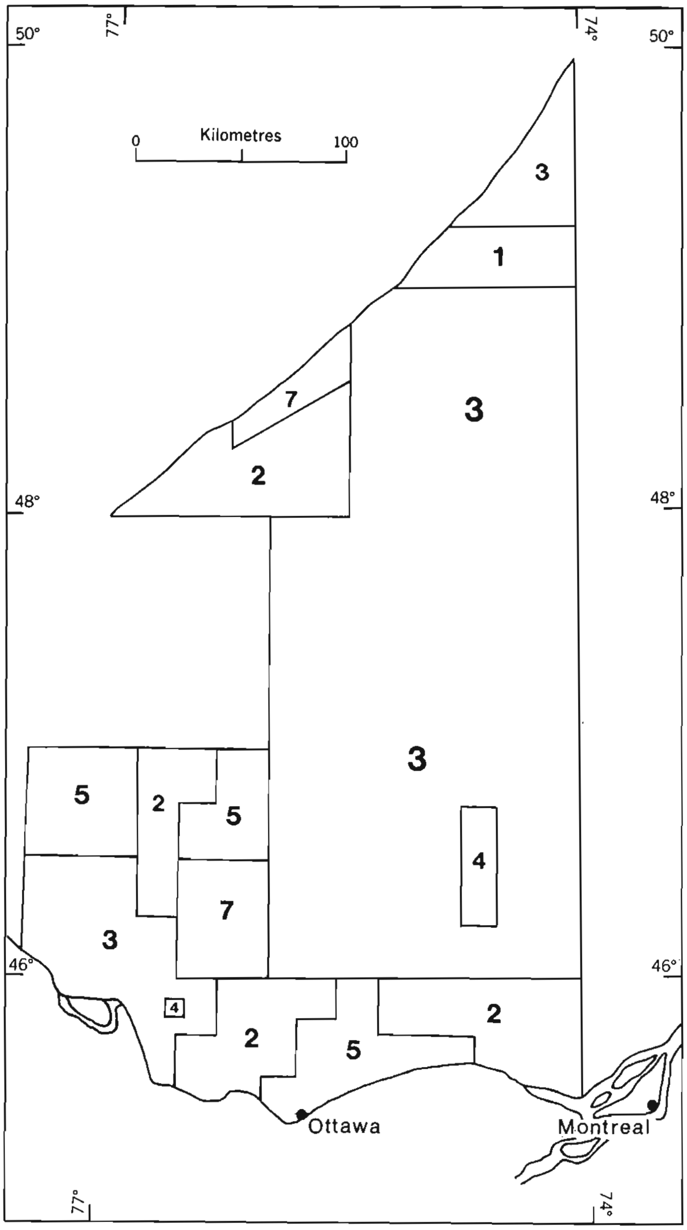

Figure 3. Reliability map of a portion of the Grenville Province of southwestern Quebec. For the significance of the different numbers, see Figure 2.

Most of the area has been mapped at a scale of 1:250 000. Mapping at a scale of 1 inch to one mile, undertaken by the Ministère des Richesses Naturelles du Québec, has covered approximately 20 000 km^2 of the eastern Grenville Province. The area encompasses many of the eastern Grenville Province tectonic units as defined by Wynne-Edwards (1972) as well as the Grenville Front tectonic zone. Only two Rb-Sr age determinations have been reported (Frith and Doig, 1975).

More than 800 thin sections and many more hand specimens were examined during the investigation. In addition, government reports, journals, and university theses were searched for relevant information, and informal discussions were held with several geologists who have worked in various parts of the eastern Grenville Province.

Granulite facies rocks

Granulite facies rocks are exposed in 7 disparate areas of the eastern Grenville Province (Fig. 4):

1. The Forteau charnockite (F, Fig. 4) near the Strait of Belle Isle contains a granulite facies mineralogy (orthopyroxene-clinopyroxene) which is probably the product of a prograde metamorphic event, but may be the result of primary crystallization from a melt (Bostock, pers. comm., 1977). The charnockite is enveloped by a zone of stable upper amphibolite facies (sillimanite-K feldspar) rocks.

2. A narrow zone of granulite facies borders the east side of the Mealy Mountains anorthosite complex (MMAC, Fig. 4). The preliminary distribution of these granulites suggests a genetic association with the emplacement of the anorthosite, however, no granulite facies rocks have been found bordering the north and west sides of the mass. A zone of upper amphibolite facies rocks parallels this granulite facies zone, on the side removed from the anorthosite. Geological control in the granulite zone and in the upper amphibolite zone is poor.

3. A small, roughly circular area of granulite facies rocks exposed northwest of the Double Mer (Fig. 4) is surrounded by lower grade rocks. The textures within the granulite mass are entirely metamorphic, with no evidence of retrogression. Near the margins a cataclastic fabric is accompanied by substantial development of secondary minerals. The secondary mineral assemblages are similar to the assemblages that characterize the surrounding lower grade rocks, suggesting that a lower grade event postdates the granulite facies metamorphism. The low grade rocks are fine grained, presumably of sedimentary or volcanic origin. The granulite mass would then represent the basement upon which these rocks were deposited.

4. Granulite facies rocks underlie about 16 000 km^2 in southern Labrador. The Winokapau-Red Wine Block in the east (WRW, Fig. 4) and the Ossokmanuan Lake Block to the west (OLB, Fig. 4) are joined together by a relatively narrow east-west trending area called the Lac Ghyvelde Complex (LGC, Fig. 4). Stevenson (1969) mapped both the Winokapau-Red Wine Block and the Lac Ghyvelde Complex as the same unit – brown weathering sillimanite gneiss. All these areas are associated with major gravity anomalies (Thomas, 1974).

The granulite facies rocks which comprise the Ossokmanuan Lake Block were first described by Fahrig (1960) and Wynne-Edwards (1961). Wynne-Edwards mentioned the presence of the assemblage sillimanite-orthopyroxene. This assemblage is rare and is restricted to very high grade metamorphic terranes. A search of the literature revealed its presence in only 13 localities, in addition to those under discussion here.

Temperatures and pressures estimated in various localities in which this assemblage is found include: Anabar Shield, USSR: 950°C, 12 kb[1] (Lutts and Kopaneva, 1968); Mozambique Belt, Uganda: 1050°C, 9 kb (Nixon et al., 1973); St. Maurice Valley, Grenville Province, Quebec: 700-800°C, 5-9 kb (Ferguson, 1974); Arunta Block, Australia: 900°C, 9 kb (Woodford, 1974; Woodford and Wilson, 1976); and the Kola Peninsula, USSR: 740°C, 10 kb (Bondarenko, 1971). In the Kola Peninsula occurrence, kyanite is the stable Al_2SiO_5 phase. The assemblages in the Mozambique Belt and in the Arunta Block contain quartz.

Marakushev and Kudryavtsev (1965) consider that the assemblage forms with increasing pressure according to the reaction:

$$\text{garnet + cordierite = sillimanite + orthopyroxene + quartz} \quad \text{(R.1)}$$

[1] 1 kb = 1 x 10^5 kPa

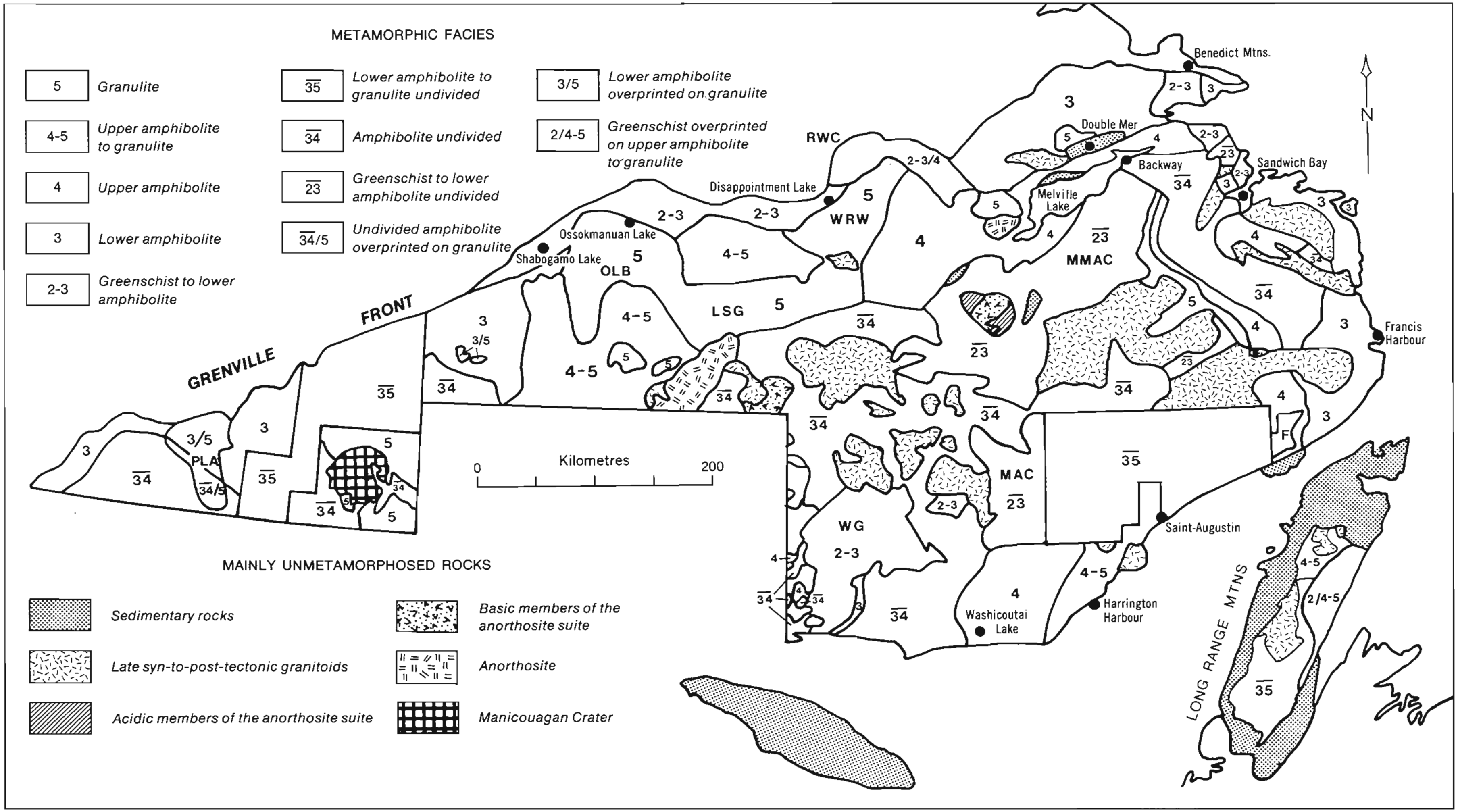

Figure 4. Sketch map showing the distribution of metamorphic facies in the eastern Grenville Province. All features referred to in the text are plotted, using the following abbreviations:

MC	MANICOUAGAN CRATER	WG	WAKEHAM GROUP
SHL	SHABOGAMO LAKE	MMAC ...	MEALY MOUNTAINS ANORTHOSITE COMPLEX
OLB......	OSSOKMANUAN LAKE BLOCK	PLA......	PAMBRUN LAKE ANORTHOSITE
LGC	LAC GHYVELDE COMPLEX	F	FORTEAU CHARNOCKITE
WRW	WINOKAPAU-RED WINE BLOCK	MAC	MECATINA ANORTHOSITE COMPLEX
RWC	RED WINE COMPLEX		

The temperatures and pressures attributed to this assemblage by workers in different areas suggests that the assemblage sillimanite-orthopyroxene-quartz is stable only under extreme metamorphic conditions. Noteworthy is the fact that sapphirine is found in many of these localities.

The assemblage kyanite-K feldspar (microperthite) present in the Ossokmanuan Lake Block is indicative of high pressure and moderate to high temperature (Carmichael, 1974). The presence of the assemblages sillimanite-orthopyroxene-quartz and kyanite-K feldspar permits some insight into the P-T contrast across the Grenville Front in this region. The rocks of the Labrador Trough are exposed immediately to the north. At distances greater than 30 km north of the front, the metamorphic grade of the Labrador Trough rocks is very low. Near the front the effects of metamorphism intensify and on the north shore of Ossokmanuan Lake the chloritoid zone of the greenschist facies is attained.

Kyanite-K feldspar rocks of the Grenville Province are exposed on the south shore of this lake, only 10 km from the chloritoid-bearing outcrops. An estimate of the P-T contrast across the lake would be 4 kb and at least 300°C. No sapphirine has yet been reported from the Ossokmanuan Lake Block.

The Winokapau-Red Wine Block[1] (WRW, Fig. 4) is approximately 110 km long and 30 km wide. Sapphirine occurs on and near Wilson Lake (Meng, 1967; Morse and Talley, 1971; Moore and Meng, 1971) and in two additional sapphirine localities which were discovered in the area during this study.

Assemblages at these two localities, inferred to have been stable at maximum temperature and pressure conditions, are: (1) sapphirine-quartz-plagioclase (An_{30})-K feldspar (microperthite)-sillimanite-cordierite(?), and (2) sapphirine-quartz-plagioclase (An_{28})-K feldspar (microperthite). Co-existing plagioclase and K feldspar (perthite host) were analyzed on the electron microprobe to calculate a temperature of equilibration using the method of Stormer (1975). The average temperature determined from sample (1) for 5 pairs was 936 ± 20°C and for 4 pairs from sample (2) was 943 ± 15°C. Both error estimates are one standard deviation about the mean. Since the K feldspar involved is clearly perthitic, the Na content of the homogenized potash feldspar grain would be higher than that obtained analytically. This would result in still higher temperature determinations than those above.

A recent study by Bohlen and Essene (1977) has shown good agreement between the above method and the magnetite-ilmenite method of temperature estimation in the granulite facies of the Adirondacks. Although a reliable assessment of the accuracy of this method must await additional studies, it is clear that these rocks have been metamorphosed to very high temperatures.

In both samples retrograde rims have developed around large sapphirine porphyroblasts, effectively isolating them from the rest of the rock. In sample (1) the rim is thin and composed of a very fine grained aggregate of sillimanite and orthopyroxene. In sample (2) the retrograde assemblage is sillimanite-orthopyroxene-biotite and the retrograde minerals are much larger in size. Both these observations suggest that sample (1) has undergone a "dry" retrogression according to the reaction:

$$\text{sapphirine + quartz = sillimanite + orthopyroxene} \qquad \text{(R.2)}$$

whereas sample (2) has adjusted to the later P-T conditions according to the reactions:

$$\text{sapphirine + quartz + K feldspar + } H_2O \quad \text{sillimanite + orthopyroxene + biotite} \qquad \text{(R.3)}$$

Both samples contain abundant quartz. Quartz was not observed in contact with sapphirine due to the presence of the reaction rims about the latter, however the texture indicates that quartz and sapphirine once formed a stable mineral assemblage.

The assemblage sillimanite-orthopyroxene was found at one locality in the Winokapau-Red Wine Block. This occurrence, unlike those in the Ossokmanuan Lake Block, is not located on a gravity anomaly.

The granulite facies rocks of the Lac Ghyvelde Complex (LGC, Fig. 4) locally contain the assemblage sillimanite-orthopyroxene, indicating that this complex also is composed of very high grade metamorphic rocks. The aeromagnetic and gravity anomaly patterns associated with this complex are similar to those associated with the Ossokmanuan Lake and Winokapau-Red Wine blocks. The relationship between these three areas is not clear.

5. Several scattered areas of granulite facies rocks have been described by Currie (1972) in the Manicouagan crater (Fig. 4) area of Quebec. Although detailed petrographic descriptions were not presented, a sillimanite-orthopyroxene assemblage was reported in the metasedimentary rocks. Currie states (page 10):

> The mineral assemblages (sic) quartz-orthoclase-garnet-sillimanite in the metasedimentary rocks, together with the general lack of hydrous minerals, and the distinctive yellowish colour in outcrop, are all characteristics of granulite grade metamorphic rocks.

A strong positive gravity anomaly is present northeast of the crater in the area underlain by a large tract of granulite facies rocks. However, anomalies are not associated with granulites northwest and southeast of the crater.

6. A small granulite facies zone was identified northwest of Lake Melville. The best exposures of the granulites are along the shore of Grand Lake (the northwest arm of Lake Melville) and the name Grand Lake Granulites has been given to these rocks. In three of four samples examined the quartz was extremely flattened. Two of the four samples contain the assemblage plagioclase-orthopyroxene-quartz, whereas the remaining two contain the assemblage plagioclase-garnet-clinopyroxene-quartz. The region is located close to the Grenville Front, which may in part explain the cataclastic nature of the rocks.

7. The Pambrun Lake anorthosite massif (PLA, Fig. 4) is very poorly known. The massif and its border complex have been severely deformed and completely recrystallized. Chown (pers. comm., 1976) has described garnet-clinopyroxene symplectites which he considers to have formed at the same time as the deformation. This would place the anorthosite in the garnet-clinopyroxene (-plagioclase) subfacies of the granulite facies according to deWaard's (1965) classification scheme.

Secondary green and blue-green hornblende have developed about the margins of the pyroxene, suggesting an incomplete adjustment to a later amphibolite facies metamorphic episode of variable intensity. Rocks to the south and west of the Pambrun Lake mass are in the green hornblende zone, whereas those to the east and northeast are in the blue-green hornblende-epidote zone. The contact between the two

[1] The nomenclature of the Winokapau-Red Wine area is presently confused. The Winokapau-Red Wine Block as used here refers to a large tract of ground extending from Lake Winokapau in the southwest to the Red Wine Mountains in the northeast. The well-known Wilson Lake sapphirine occurrences are located near the southern limit of the block. The northern end of the block has recently been called the Ptarmigan Complex (Emslie et al., 1978). However, this name conflicts with an established formal name and will henceforth be known as Ptarmigan Lake Complex.

zones passes diagonally across the anorthosite. The anorthosite and associated rocks are associated with a marked gravity high.

Transitional Amphibolite — Granulite Facies

Four areas, classified as transitional from amphibolite to granulite, are characterized by both upper amphibolite facies and granulite facies assemblages which are intimately and inseparably associated, or are separable but not on the scale of Figure 4.

All of the area is in the sillimanite-K feldspar grade (or higher grade) of regional metamorphism. The assemblage orthopyroxene-plagioclase is present in some of the 'mafic' units, whereas the assemblage hornblende-garnet is present in others. Hornblende-garnet assemblages generally predominate over orthopyroxene-bearing assemblages. It is in these areas that the comparatively rare assemblage garnet-cordierite is found. Sillimanite grains in the rocks of this zone are prismatic and well formed.

1) The northern quarter of the Long Range Mountains of Newfoundland (Fig. 4) has been placed in this facies classification, based exclusively on the work of Bostock et al. (in prep.) who found only "traces of granulite facies metamorphism", and attributed this to the fact that the rocks "only barely reached granulite facies conditions". There is no spatial relationship between the distribution of the granulite facies rocks and the metagabbro masses of the region, which he considers to be a part of the anorthosite suite of intrusions.

Information on the remainder of this area is very limited. Bostock and Cumming (1973) described a small area around St. Paul's Inlet in the extreme southwestern corner of the Long Range underlain by rocks similar to those of the northern Long Range. These rocks are predominantly gneisses in the upper amphibolite facies that locally contain traces of orthopyroxene. The aeromagnetic pattern of the high grade rocks in the St. Paul's area and in the northern portions of the Long Range are matched throughout the central, unmapped portions of the area. It is therefore probable that the entire western half of the Long Range is composed of rocks of transitional amphibolite to granulite facies.

Intrusive into the Long Range gneisses are a number of megacrystic granite plutons. Andalusite occurs at this location on the margins of these masses (Bostock, pers. comm., 1976). The intrusions, therefore, were relatively shallow and could only have been emplaced after a period of considerable erosion of the high grade area. Intrusive into these granite plutons is a swarm of northeast-trending diabase dykes. In the western portion of the northern Long Range the dykes are not metamorphosed; however, towards the east they show increasing development of secondary minerals, particularly epidote. This is a consequence of a Paleozoic greenschist facies overprint. The boundary between the affected and unaffected areas is completely gradational. A subtle difference in the grain of the aeromagnetic pattern in the north appears to be correlated with the distribution of the greenschist facies overprint and this has been used as a basis for approximating the boundary of the overprinted greenschist facies terrane farther south in the poorly known parts of the Long Range Mountains.

2) Davies (1963, 1965a, b) has described the mineralogy of the gneiss across the Gulf of St. Lawrence in the Harrington Harbour-St. Augustin area (Fig. 4) and considers that "the mineralogy is that of the granulite and upper amphibolite facies" (Davies, 1965a, p. 2). In pelitic rocks the common assemblage is biotite-garnet-sillimanite(prismatic)-graphite. Also present is a distinctive greenish pyroxene-plagioclase gneiss unit. The green colour is due to the presence of green plagioclase. No mention was made of the nature of the pyroxenes. Muscovite is restricted to fracture surfaces and is probably of secondary origin. Cordierite was not reported in this unit.

3) The third area classified as transitional upper amphibolite to granulite facies lies immediately north of the Lac Ghyvelde Complex. A common assemblage in this area is biotite-perthite-sillimanite (prismatic) ± garnet. In one locality highly altered cordierite (?) coexists with garnet. One occurrence of orthopyroxene was noted. In brief, these rocks contain many characteristics of the regions described above by Bostock and by Davies, and on this basis have been classified as transitional upper amphibolite to granulite.

4) The fourth area classified as transitional upper amphibolite to granulite facies lies south of the Ossokmanuan Lake Block. The mineral assemblages present resemble those of the area just described. Biotite-perthite-sillimanite (prismatic) assemblages are very common. Orthopyroxene is present in some of the basic rocks.

The last two areas of transitional upper amphibolite to granulite facies rocks, together with the Ossokmanuan Lake Block, the Lac Ghyvelde Complex, and the Winokapau-Red Wine Block, which belong to the granulite facies, collectively form an immense area of very high grade metamorphic rocks. The age of the metamorphic episode responsible for the assemblages in these rocks is unknown.

Upper Amphibolite Facies

Upper amphibolite facies rocks are located in five separate regions of the eastern Grenville Province.

1) Upper amphibolite facies rocks are found immediately east of the Winokapau-Red Wine Block. The relationship between these rocks and those of the Winokapau-Red Wine and Ossokmanuan blocks is obscure. The correlation between the high metamorphic grade and distinctive aeromagnetic and gravity signatures for the Winokapau-Red Wine and Ossokmanuan granulite terranes suggests that both blocks are exotic and fault-bounded on all sides. If this is the case, no spatial relationship between the upper amphibolite and granulite facies rocks should be expected. However, Figure 4 shows that this is the largest upper amphibolite facies terrane identified in the eastern Grenville Province. Its position adjacent to the largest granulite terrane in the eastern Grenville Province certainly suggests a possible correlation between the two terranes.

The upper amphibolite facies area contains large tracts of sillimanite-perthite-bearing rocks, but only rare orthopyroxene in spite of the presence of numerous mafic horizons. Prograde muscovite is lacking throughout, but "secondary" muscovite was found in the southeastern portion of the outcrop area near the Mealy Mountains.

These rocks pass southwards into an area of undivided amphibolite facies rocks. The boundary between the two is arbitrary. South of the boundary no prograde muscovite or sillimanite-K feldspar assemblages were noted. This is possibly related to a facies change in the original sediments.

2) A small region of upper amphibolite facies rocks occurs in the Sandwich Bay area of eastern Labrador (Fig. 4). The diagnostic assemblage here is sillimanite-K feldspar (untwinned). One sample containing orthopyroxene with clinopyroxene may be of igneous origin. The relationship of this small region of high grade rocks to the large surrounding area of lower grade rocks is not known.

3) Upper amphibolite facies rocks exposed north of the Gulf of St. Lawrence in the Washicoutai Lake area (WL, Fig. 4) (Bourne et al., 1978) contain sillimanite-K feldspar and rare

garnet-cordierite assemblages. All muscovite encountered in quartzofeldspathic gneiss is secondary. These rocks grade into the transitional upper amphibolite to granulite facies terrane in the Harrington Harbour-St. Augustin area to the east. As the boundary between the two terranes is approached from the west, the colour of the quartzofeldspathic gneiss changes abruptly from pink to beige, and the gneiss becomes crumbly and deeply weathered. To the northwest this terrane grades into a zone of undivided amphibolite rocks. It is probable that these rocks are also in the upper amphibolite facies since migmatites are found in many places; however, the bulk composition is not favourable for the formation of either sillimanite-potash feldspar or muscovite-quartz assemblages.

The remaining two regions underlain by upper amphibolite facies rocks, described earlier, occur east of the Mealy Mountains anorthosite complex and around the Forteau charnockite.

Lower and Middle Amphibolite Facies

Recognizeable lower and/or middle amphibolite facies rocks are confined to two areas in the eastern Grenville Province.

1) Rocks containing muscovite-quartz assemblages are exposed along the Atlantic coast of Labrador and for considerable distances inland. They also contain abundant stable epidote, blue-green hornblende, sphene, microcline and pleochroic green to black biotite. The plagioclase composition is generally within the range An_8-An_{18}. Plagioclase in contact with K feldspar commonly has thin albite (An_{0-7}) overgrowths.

Wardle (1977; pers. comm. 1976) described a small portion of this area near Francis Harbour, originally mapped by Eade (1962). A large proportion of the rocks are migmatites and in general all rocks appear to have undergone high grade metamorphism. However, only the comparatively low grade mineral assemblages just described were found in thin section. It is concluded that either the terrane was completely retrograded upon cooling from the maximum P-T conditions during metamorphism, or that a second pervasive metamorphic episode has completely destroyed all mineralogical record of the first.

2) Lower amphibolite facies rocks are also exposed over a large region west of the Ossokmanuan Lake Block. The easternmost limit of these rocks has been chosen as the outcrop area of the Gagnon Group metasedimentary rocks shown on the Opocopa Lake map (Jackson, 1976). Details of the metamorphic grade of the Gagnon Group rocks in the Opocopa area are sparse as no samples were available for study. However, on the legend accompanying this map two units were defined by the following useful metamorphic mineral assemblages:

quartz-biotite-kyanite schist and gneiss in which plagioclase muscovite, garnet and graphite are common, and

quartz-feldspar-muscovite-biotite-kyanite schist and gneiss in which garnet and graphite are common.

Both map units presumably contain the assemblage muscovite-quartz-kyanite-garnet-biotite, which is diagnostic of high pressure middle amphibolite facies metamorphism (Hess, 1969). The distribution of these units defines the extent of this metamorphism. Jackson's map shows these units throughout the western part of the Opocopa sheet, with the exception of the southwest corner, where the rocks may be of higher grade, or bulk compositions of the units may be significantly different.

Rocks of similar metamorphic grade in the Mt. Wright area to the west include biotite-muscovite gneiss and minor amounts of muscovite and kyanite gneiss (Duffell and Roach, 1959). Some small areas of Archean granulite facies rocks exposed in the Mt. Wright area (similar to those immediately north of the Grenville Front) have been overprinted by this later middle amphibolite facies metamorphism (Clarke, 1968). Two such areas are shown on Figure 4.

The Opocopa-Mt. Wright middle amphibolite facies rocks are separated from their apparent equivalents to the southwest and west by an area in which no mapping has yet been undertaken.

Some information from the Pambrun Lake area rocks is contained in a series of preliminary reports (Bérard, 1964; Chown, 1964, 1971a, b; Chown and Hashimoto, 1965; Chown and Laurin, 1970; Murtaugh, 1965). The minerals in the Pambrun Lake area include blue-green hornblende, abundant epidote and rare garnet. These rocks grade southwards into a green hornblende zone which has been shown on Figure 4 as amphibolite facies (undivided). Kyanite has not been reported from this area, although it is present north of the Grenville Front in the Otish Mountains (Chown, pers. comm., 1976).

Amphibolite Facies (Undivided)

Undivided amphibolite facies rocks, all of which lack useful index minerals and/or mineral assemblages, are distributed in three large areas.

1) The largest of these areas is located between the Lac Ghyvelde Complex (LGC, Fig. 4) and the Washicoutai Lake area (WL, Fig. 4). The definitive assemblage in this region is green hornblende-garnet. Only one sillimanite-K feldspar assemblage and no prograde muscovite-quartz assemblages have been reported. In the eastern portions of this terrane relatively large flakes of "secondary" muscovite are common. This muscovite is possibly related to a later metamorphic overprint. It is equally possible, however, that the metamorphic grade was simply lower in this region than in the rest of the undivided amphibolite area, and that the muscovite is primary. The plagioclase composition is generally An_{15}, with rims of An_{0-5}. K feldspar generally shows microcline grid twinning. Green biotite is ubiquitous.

2) A large tract of undivided amphibolite facies rocks east of the Mealy Mountains extends from the vicinity of the Backway (B, Fig. 4) southeast towards Francis Harbour (Fig. 4). Texturally these rocks are identical to the lower amphibolite facies rocks which outcrop to the east along the Atlantic coast. This region was classified as undivided amphibolite facies because the composition of the stable plagioclase is always greater than An_{17}, the hornblende is green in colour (as opposed to blue-green near the coast), and neither muscovite-quartz nor sillimanite-K feldspar assemblages were noted. Cummingtonite was found in two localities, both in plagioclase-free quartzites. Secondary chlorite and sericite are common throughout this zone. In common with the lower amphibolite facies rocks of the Francis Harbour area, the K feldspar phase displays excellent microcline twin patterns; epidote, however, is absent.

3) A large undivided amphibolite facies zone extends from the vicinity of the Ossokmanuan Lake Block to the southwest, more or less parallel to the Grenville Front. Metamorphic data in this zone are virtually nonexistent and the zone boundaries shown in Figure 4 are gross approximations. A prograde relationship is probable between rocks closer to the Grenville Front, which contain blue-green hornblende, and those of this zone, which contain green hornblende.

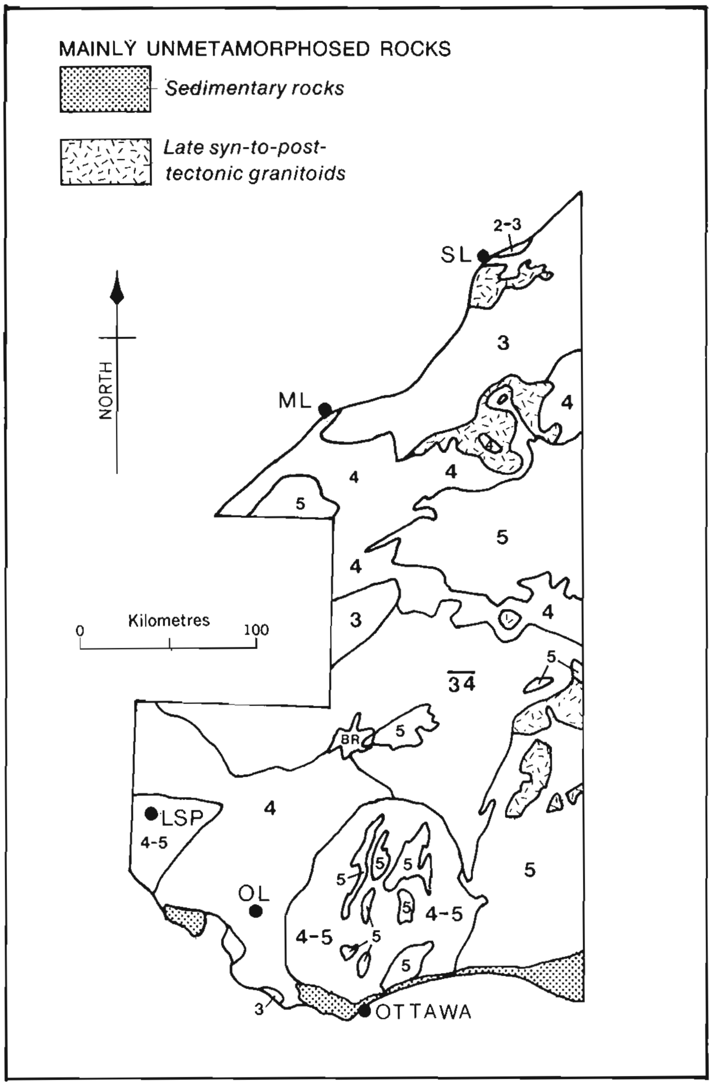

Figure 5. Sketch map showing the distribution of metamorphic facies in a portion of the southwestern Grenville Province. Geographical localities referred to in the text are indicated by the following abbreviations:

SL ..... SURPRISE LAKE
ML ..... MÉGISCANE LAKE
BR ..... BASKATONG RESERVOIR
LSP .... LAC ST-PATRICE
OL ..... OTTER LAKE

Metamorphic facies keyed as in Figure 4.

Transitional Greenschist to Lower Amphibolite Facies

Two large areas of rocks metamorphosed to conditions close to the greenschist-amphibolite facies boundary were recognized. These are: the Wakeham Group area (WG, Fig. 4) in the Havre St. Pierre area of the eastern Grenville Province, and a long narrow zone parallel to the Grenville Front. The classification of these rocks as either upper greenschist or lower amphibolite facies was found to be impossible due to the lack of characteristic index minerals such as chloritoid and staurolite.

Wakeham Group Area

The first mention of the rocks of the Wakeham Group as upper greenschist to lower amphibolite facies was made by Sharma (1971), although detailed reports by some earlier workers (Claveau, 1949; Cooper, 1957; Grenier, 1957) indicated that low grade rocks were widespread throughout this group.

The Wakeham Group consists of a thick succession of metasandstone with minor amounts of interbedded meta-arkose, metaconglomerate, and metacarbonate. Arkose and conglomerate are most common in the southeast corner of the map area. Predominantly acidic metavolcanic rocks associated with the sandstone throughout a large area north of 51°N, are considered to be roughly contemporaneous with the Wakeham Group rocks and have been metamorphosed to the same grade (Bourne et al., 1977). All the aforementioned units have been intruded by gabbroic sills and dykes which have also been metamorphosed.

The western boundary of the Wakeham Group rocks is not exposed in the area. The southeastern and northwestern boundaries are faults. Throughout the rest of the area the boundary is presumably a deformed unconformity which in some localities is the locus of later granite intrusions.

The metamorphic grade of the rocks is close to the greenschist-lower amphibolite facies boundary. Chloritoid has not been positively identified in this area; however, Grenier (1957, plate Xb) published a photograph of an unidentified porphyroblast which greatly resembles this mineral.

Prograde chlorite is present throughout the region, as are sericite (muscovite-quartz), epidote, rutile and sphene. The assemblages biotite-epidote-ankerite-quartz and biotite-epidote-calcite-chlorite, both of which are found in the area, indicate a relatively low temperature of metamorphism (Carmichael, pers. comm., 1973).

Within 8 km of the southeastern limit of the Wakeham Group, the metamorphic grade increases to lower amphibolite facies. Two staurolite-bearing assemblages have been reported: staurolite-garnet (Blais, 1955) and garnet-staurolite-kyanite-"micas" (McPhee, 1961). If both biotite and muscovite are present in the second assemblage, as implied, the mineral assemblage would have formed under moderately high pressure, middle amphibolite facies conditions.

Southeast of the fault forming the southeastern limit of the group, a large area of "augen gneiss" has been delimited by several authors (Bassaget, 1970a, b, 1972; Blais, 1955; McPhee, 1959). Blais emphasized that the metamorphic grade of the augen gneiss is higher than the grade of the adjacent Wakeham Group rocks west and northwest of the fault. The gneiss has therefore been tentatively assigned to the undivided amphibolite facies.

Grenville Front Zone

The part of the Grenville Front zone described below extends from the south end of the Labrador Trough to the Atlantic Ocean, a distance of approximately 600 km. Along this zone relatively low grade rocks are common. Some of these are clearly a transitional phase between lower grade rocks to the northwest and the high grade rocks typical of most of the Grenville Province to the southeast. The transition between the Superior and Grenville provinces in the Surprise Lake area, southwest of Chibougamau (Fig. 5), is an example of this feature. In other areas these low grade rocks appear to be in fault contact with high grade Grenville Province rocks and the original relationship between the two is obscure.

The available data in this zone are concentrated in the Benedict Mountains area (Stevenson, 1970) and in the Disappointment Lake area (Stevenson, 1969).

All the rocks in the area south of the Benedict Mountains are fine to very fine grained. The composition of the plagioclase co-existing with epidote is commonly in the An_6 to An_{17} range, but many plagioclase grains have marginal overgrowths of composition An_{0-2}. The biotite in these rocks shows deep green to black pleochroism. Chlorite, probably Mg-rich, is present as a stable phase.

A large shear zone extending 100 km west from the Atlantic coast separates the Benedict Mountains from the greenschist and lower amphibolite facies rocks to the south. The zone outcrops westward towards the Disappointment Lake area (Fig. 4) and also extends southeast across Lake Melville to the Sandwich Bay area, where it appears to grade into rocks of higher (lower amphibolite) metamorphic grade. South of Sandwich Bay the grain size of the rocks is coarser, and hornblende is blue-green to green, as opposed to pale green (actinolite?).

In the area of transitional greenschist to lower amphibolite facies rocks south of the Benedict Mountains, is a swarm of gabbroic dykes known as the Michael Lake gabbro. The length of this swarm is about 275 km. The dykes have been metamorphosed by the same event which affected the finer grained rocks discussed above. A Rb-Sr whole-rock isochron age determination of 1488 ± 107 Ma ($\lambda=1.39\times10^{-11}$yr) (Fahrig and Larochelle, 1972) is considered to represent the age of emplacement of the dykes. Both Grasty et al. (1969) and Taylor (1971) have suggested that the Grenville Front passes through the middle of this area.

The low grade rocks of the Disappointment Lake area (DL, Fig. 4) are immediately west of the very high grade rocks of the Winokapau-Red Wine Block. Plagioclase composition ranges from less than An_7 to more than An_{15}. Epidote and muscovite are common. Farther north, in the Red Wine Mountains, Currie et al. (1975, p. 277) attributed low grade mineral assemblages to a "late retrograde metamorphism to upper greenschist facies, as suggested by the muscovite-biotite-epidote assemblages in the Wapustan Gneiss" and on the basis of field evidence postulated an earlier, amphibolite facies metamorphic event. No trace of this event has been recognized in the Disappointment Lake area. The area considered by Currie to be polymetamorphic is shown on Figure 4.

Greenschist and/or Lower Amphibolite Facies

The Mealy Mountains anorthosite complex and associated rocks in south-central Labrador have been metamorphosed to greenschist and/or amphibolite facies. Emslie (pers. comm., 1977) indicated that all rock types in this complex have been deformed and metamorphosed, including the dykes that intrude them. Most members of the suite show evidence of secondary replacement of the primary pyroxenes. Radiating aggregates of pale green amphibole (actinolite) or pale green chlorite commonly surround the mafic minerals. Near the core of the complex the igneous rocks appear to be unmetamorphosed.

It is probable that the metamorphic episode which was responsible for the above-mentioned mineral development in the anorthosite complex also generated the extensive zone of secondary muscovite in gneiss throughout the eastern Grenville Province, as well as the mineral assemblages found in the two transitional greenschist to lower amphibolite facies terranes described above. The main metamorphic event which generated the upper amphibolite and undivided amphibolite mineral assemblages on the west side of the Mealy Mountains was likely pre-anorthosite.

Shock Metamorphism

The only area in the eastern Grenville Province in which features attributed to shock metamorphism have been reported is the Manicouagan Crater area of central Quebec, a large circular structure outlined by Manicouagan and Mouchalagane lakes. The structure comprises a marginal rim of early Mesozoic volcanic rocks, described as larvikites and trachy-andesites, which enclose a core of shock deformed gneiss which is the equivalent of the granulite facies gneiss surrounding the structure. The core gneiss contains one or more of the following features: vesicular glass, deformation lamellae, maskelynite and optically anomalous pyroxene and quartz. The petrography of the rocks in the crater has been described in considerable detail by Currie (1972).

SOUTHWESTERN GRENVILLE PROVINCE

The part of the southwestern Grenville Province under discussion (Fig. 5) underlies approximately 70 000 km^2. Parts of the area have been mapped by the Quebec Department of Mines at a scale of 1 inch to 1 mile. Interpretative regional compilations have been made by Wynne-Edwards (1969), Wynne-Edwards et al. (1966) of the Mont Laurier-Kempt Lake area, and by Baer et al. (1978) of the Precambrian terrane in this part of Quebec and adjacent Ontario.

The metamorphic interpretation presented here is derived principally from published reports. About three hundred thin sections representing rocks between Otter Lake (OL, Fig. 5) and the Baskatong Reservoir (BR, Fig. 5) were examined.

Granulite and Transitional Upper Amphibolite to Granulite Facies

Granulite facies and transitional granulite to upper amphibolite rocks underlie almost one third of the area shown in Figure 5.

A broad granulite facies zone was delimited by Wynne-Edwards et al. (1966) in the southeastern corner of the area. This zone contains metamorphosed rocks of the Grenville Supergroup of sedimentary origin and the anorthosite suite of intrusives. All these rocks have been transformed into a complex of charnockitic gneiss, leuco- and mafic granulites, and two-pyroxene gneiss. The charnockitic gneiss, the most widespread and voluminous of the four units, was described as medium grained, and characterized by brown-weathering green perthitic feldspar, biotite, hypersthene, and minor hornblende, clinopyroxene or garnet (Wynne-Edwards et al., 1966). Brown hornblende is the most common mafic mineral in the gneiss. The assemblage garnet-cordierite was observed in only one specimen.

Katz (1973, p. 110) presented a detailed account of the mineral assemblages and inferred metamorphic grade of a relatively small area of the leucocratic granulite unit of Wynne-Edwards et al. (1966). Two thirds of the rocks sampled were assigned to the granulite facies, the remainder to the amphibolite facies. This suggests that the transitional upper amphibolite to granulite facies classification is more appropriate for much of the terrane originally classified as granulite facies by Wynne-Edwards. The granulite facies classification has been retained for the "green rock complexes". The transitional zone has been extended to the west to include four known kornerupine localities north of Ottawa (Fig. 5) and the hornblende-orthopyroxene rocks in the Otter Lake area (Kretz, 1971, p. 35).

A relatively small area of granulite facies rocks was identified by Wynne-Edwards near the Baskatong Reservoir (BR, Fig. 5). A more detailed study of this area has been carried out by Jacoby (1975) who concluded that the rocks "belong to the upper amphibolite facies or locally to the granulite faces" (p. 43). They have been classified in this paper as transitional amphibolite to granulite facies.

A large area of granulite facies rocks, centred 150 km northeast of the Baskatong Reservoir, was mapped by Wynne-Edwards et al. (1966). In accordance with Wynne-Edwards, they have been classified as granulite facies, although the transitional granulite to amphibolite facies classification may be more appropriate. The pyroxene-gneiss of this area extend into the adjoining Gouin Reservoir area mapped by Laurin (1965) who has presented mineralogical descriptions that clearly indicate that the gneisses are in or are close to the granulite facies.

Upper Amphibolite and Undivided Amphibolite Facies

Upper amphibolite facies rocks underlie a substantial portion of the remainder of this part of the southwestern Grenville Province. This terrane is located between the two large granulite zones defined by Wynne-Edwards (Fig. 5). Wynne-Edwards (1969, p. 168) considered that most of the rocks in this area should be classified as middle amphibolite facies, based chiefly on the common occurrence of the mineral assemblages quartz-almandine-muscovite and quartz-almandine-K feldspar. Recent work in the Gracefield-Maniwaki region north of Ottawa, however, indicates that this area contains no prograde muscovite. Garnet-cordierite assemblages and sillimanite-K feldspar assemblages are widespread, but not common, and orthopyroxene is rare. Rare calcareous marker horizons (skarns) are composed of mineral assemblages that include diopside-wollastonite. Amphibolite layers contain green hornblende and commonly small amounts of garnet.

It is likely that much of the terrane classified by Wynne-Edwards as middle amphibolite facies should be reclassified as upper amphibolite facies. Those areas classified by Wynne-Edwards as upper amphibolite facies, or which are known on the basis of thin section examination to be in the upper amphibolite facies, have been so indicated (Fig. 5). Areas classified by Wynne-Edwards as middle amphibolite facies, and for which there are no additional data, have been assigned to the undivided amphibolite facies.

Upper amphibolite to granulite facies rocks in the Lac St. Patrice area have been described by Katz (1969, 1976) who mapped an orthopyroxene isograd in the basic rocks of the region. It should be noted that his interpretation of granulite facies rocks in the northwest corner of the Lac St. Patrice map area is not in accord with that of Rive (1970, 1977), who has mapped several adjacent map areas. An interim classification of transitional upper amphibolite to granulite facies has been assigned here.

The Grenville Front Transition Zone — Lower and Middle Amphibolite Facies

The boundary between the Grenville Province and the Superior Province is a transition zone of variable width and generally of lower amphibolite grade. This zone ranges in width from 3 to 5 km in the Surprise Lake area (SL, Fig. 5) near Chibougamau (Deland, 1955) to 10 km in the Mégiscane Lake area (ML, Fig. 5) (Charbonneau, 1973). The geology has been studied in considerable detail, particularly in the Chibougamau area (Neale, 1959; Imbault, 1959; Holmes, 1959; Lyall, 1959; Gilbert, 1959; Deland and Grenier, 1959). The transition zone in the Surprise Lake area has been studied by Norman (1936) and by Deland (1955). The low grade rocks of the Superior Province grade eastward into higher grade, garnetiferous equivalents in the Grenville Province. Evidence of minor faulting is present, but there is no major transverse fault in this region. The emplacement of granite into this transition zone is apparently quite common (Deland, 1955; Wegria and Bertolus, 1976).

Deland (1955) has described the common mineral assemblages from Archean basalt units of the Superior Province as colourless amphibole-plagioclase-epidote-chlorite and minor quartz. In the transition zone this assemblage changes to blue-green hornblende-epidote-chlorite and minor quartz. In the Grenville Province the common assemblage is green hornblende-garnet-epidote-plagioclase-biotite and minor quartz.

A similar transition zone has been identified in the Mégiscane Lake area by Charbonneau (1973) and by Charre (1975) who agree on the width of the zone (7-15 km) but not on its location. The problem is compounded by the lack of bulk compositions that produce useful diagnostic metamorphic assemblages. Sixty-five thin sections prepared by Charbonneau were re-examined and the following observations made:

epidote is confined to the northern two thirds of Charbonneau's transition zone;

green hornblende is found throughout the transition zone, however, pale green or blue-green hornblendes are confined to the northern two thirds of the transition zone; and

muscovite appears to be stable with quartz throughout the transition zone.

Within this transition zone the position of the Grenville Front is completely arbitrary.

In spite of the large amount of work which has been done in the Chibougamau area, few useful mineral assemblages have been reported from the transition zone. Neale (1959, p. 19) reported the assemblage staurolite-kyanite-biotite-garnet-muscovite-quartz 50 km southeast of Chibougamau. This assemblage is stable under moderate to high pressure conditions (Hess, 1969). Cimon (pers. comm., 1975) has found two other kyanite occurrences in the same general area. Laurin (1965) reported the assemblage biotite-garnet-kyanite along the southern margin of the transition zone determined by Charbonneau in the Mégiscane Lake area.

Other Minerals

Low grade minerals of secondary origin are reported with increasing frequency from accessible parts of the Grenville Province. Hogarth (pers. comm., 1976) has found several occurrences of prehnite along joint surfaces north of Ottawa, and the writer has made similar discoveries in the area northeast of Ottawa. Pumpellyite also occurs in this general area. The mineral chabazite has been found in an impure marble northwest of Otter Lake. It is probable that these minerals are common but have been overlooked. Their significance cannot be assessed until their distribution has been accurately established.

DISCUSSION

The concept of the Grenville Province as a terrane of predominantly high grade metamorphism is essentially correct. Recognizable greenschist, transitional greenschist-amphibolite and lower amphibolite facies rocks collectively underlie less than 5 per cent of the Grenville Province. The remainder is composed of upper amphibolite facies, undivided amphibolite and granulite facies rocks, and some unmetamorphosed post-Grenvillian intrusions. Of these, the upper amphibolite facies and undivided amphibolite facies are predominant. Most of the undivided amphibolite facies terrane shown in Figures 2 and 3 is probably in the upper amphibolite facies. This cannot be confirmed, however, because the bulk composition is not conducive to the development of the diagnostic sillimanite-K feldspar assemblage.

Sillimanite is the most common aluminosilicate polymorph in the Grenville Province, reflecting the predominance of rocks in the upper amphibolite facies. Sillimanite at this grade is stable at pressures ranging from 1 to 8 kb. Andalusite, which is virtually absent from the Grenville Province, would only be stable at upper amphibolite facies grade under extremely low pressures of metamorphism; kyanite would be stable only under extremely high pressures.

Kyanite has been found in two major geological settings in the Grenville Province. The first is in the rocks of the Wakeham Group which are of lower amphibolite facies grade (staurolite zone). The second is in a broad zone parallel to the Grenville Front that extends as much as 60 km into the Grenville Province itself. Commonly the kyanite co-exists with potash feldspar within this zone, particularly in the Sudbury-Temiskaming segment of Ontario. This mineral pair forms under high pressure and moderate to high temperature conditions (Carmichael, 1974; Hess, 1969). The restriction of kyanite largely to this zone suggests that the Grenville Front may have developed under pressures higher than those prevailing in the rest of the Grenville Province. It remains to be demonstrated that the rocks within this zone have been metamorphosed to approximately the same temperatures as the rocks in the remainder of the Grenville Province to the southeast. If such were not the case then the kyanite-sillimanite transition could also be explained by a temperature increase at constant pressure.

In a classification of metamorphosed basic rocks proposed by Froese (1973, p. 61), the mineral assemblage cummingtonite-plagioclase is diagnostic of low pressure regional metamorphic terranes, whereas the assemblage hornblende-garnet is considered diagnostic of high pressure regions. The only cummingtonite-plagioclase assemblages in the Grenville Province are in the Whetstone Lake-Fishtail Lake area (Carmichael, pers. comm., 1977; Lal and Moorhouse, 1969, p. 147). It is significant that this is also one of the very few areas in the Grenville Province in which andalusite has been reported. The two cummingtonite-bearing rocks observed by the writer during this investigation were plagioclase-free quartzites. The assemblage hornblende-garnet is very common throughout the Grenville Province.

In the eastern Grenville Province there is a sporadic correlation between the "topography" of the aeromagnetic maps and the distribution of high grade rocks, particularly granulite facies rocks. An irregular and variable magnetic pattern is associated with the largest of the granulite terranes of the Winokapau-Red Wine Block-Ossokmanuan Lake Block-Lac Ghyvelde Complex and also with the transitional amphibolite/granulite terrane of the Long Range Mountains of Newfoundland. This correlation is probably related to later retrograde metamorphism of the granulite facies rocks with replacement of some of the orthopyroxene and/or clinopyroxene by green hornblende, plus some iron oxide(s). Some granulite terranes, i.e. the Double Mer charnockite, do not show this relationship.

Anorthosites and associated charnockitic gneiss have posed special problems in the compilation of the metamorphic data. The anorthosites of eastern Canada occur in the Grenville Province and in Labrador north of the Grenville Province (Emslie, 1974, Fig. 1, p. 527). This distribution indicates that the generation of anorthosite and anorthosite-related rocks is not dependent on a process that is unique to the Grenville Province. The Grenville Province anorthosites have been metamorphosed and deformed during the Grenvillian Orogeny. The Labrador anorthosites, by contrast, are neither metamorphosed nor deformed, and are therefore more suitable for study of anorthosite genesis. Adamellites show a crude spatial association with the Labrador anorthosites. They commonly occur as marginal phases around the anorthosite masses, but many adamellites occur as discrete intrusions far removed from any known anorthosite. In addition, there are many localities in Labrador where the anorthosite massifs intrude gneiss of various metamorphic grades and are not separated from the gneiss by adamellite envelopes. The adamellites are fresh, massive, commonly porphyritic rocks and pea-green in colour. The primary mafic mineralogy consists of orthopyroxene-clinopyroxene or fayalite-clinopyroxene assemblages.

Not all anorthosites in the Grenville Province are associated with granulite terranes. Those that are not are most common in the eastern Grenville Province (Mealy Mountains (MMAC, Fig. 4) and Romaine River massifs (Sharma et al., 1978)). Those in the southwestern Grenville Province are in large part associated with granulite facies metamorphic rocks which have been variously described as monzonites, hypersthene syenites, charnockites (senso stricto) and charnockitic gneiss. Many of these rocks may be the metamorphosed equivalents of the adamellite rocks of Labrador – that is to say, they may have originally been essentially anhydrous.

Kehlenbeck (1972) who studied the variation in texture across the Lac St. Jean anorthosite, noted that the change from slightly deformed ophitic textures to gneissic streaky textures is not necessarily accompanied by a change in primary mineralogy (i.e. pyroxene) of the anorthosite (Plates 2 and 4 of Kehlenbeck, 1972). It is possible that a similar process operates in the case of the adamellite rocks associated with Grenvillian anorthosites – the adamellite rocks may acquire a metamorphic texture and yet retain primary orthopyroxene and clinopyroxene. The result would be a granulite facies charnockitic gneiss.

Much has been written about the "dry" nature of anorthosites. However, the average total water content of 62 anorthosites from Labrador (Emslie, in press) is 0.48 ± 0.20 per cent[1], whereas the average total water content of 33 adamellites from Labrador is 0.54 ± 0.28 per cent. These values show that adamellites and anorthosites are of approximately equal "dryness". It is suggested therefore that the lack of water in the adamellites would inhibit the development of retrograde hydrous minerals. The primary orthopyroxene and clinopyroxene of the adamellites may be preserved when the minerals are crushed, and the rock may take on the appearance of a "charnockitic gneiss". If this hypothesis concerning the possible adamellite parenthood of the charnockitic gneiss units is valid, the orthopyroxene and clinopyroxene of the charnockite must be mainly primary. The only new minerals formed were hornblende and biotite, by retrogression of the primary pyroxenes. The charnockitic gneiss was thus in part retrograded to the same metamorphic grade as the surrounding gneiss (undivided amphibolite facies).

The Grenville adamellites and associated anorthosites by analogy with the Labrador anorthosites and adamellites, may have been emplaced at intermediate crustal levels in Grenville Province gneisses.

The lack of granulite gneiss associated with the anorthosites in the eastern Grenville Province is probably related to the lack of adamellite intrusions. There were therefore no primary pyroxene-bearing rocks to be converted into "charnockitic gneisses". In the subsequent low grade metamorphism of the Grenvillian Orogeny the anorthosite was partly recrystallized, the Wakeham Group rocks were

[1] Errors quoted are one standard deviation about the mean.

raised to greenschist or lower amphibolite facies, and the earlier higher grade gneiss into which the anorthosite was originally emplaced was retrograded.

In the southwestern Grenville Province, where the Grenvillian Orogeny was apparently of greater intensity, upper amphibolite and granulite facies assemblages formed in the pre-existing gneiss. This metamorphism and the textural conversion of adamellite complexes are considered responsible for the large granulite facies terranes in the Grenville Province.

Others, (Rondot, pers. comm., 1977) prefer a deep-level intrusion hypothesis. According to this hypothesis the emplacement of hot anorthosite into rocks which were already undergoing high grade (amphibolite to granulite facies) metamorphism, raised all the rocks in the vicinity of the intrusion to granulite facies grade. The retrograde mineral development described earlier might be related to cooling following the high grade metamorphic episode, to a later metamorphism, or both. According to this scheme, the charnockitic gneiss would be portrayed as granulite facies with or without an amphibolite facies overprint.

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STRATIGRAPHIE ET METAMORPHISMES DE LA REGION DU SAINT-MAURICE

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Résumé

Quatre nouveaux groupes sont reconnus dans la région du Saint-Maurice; ce sont: le complexe de Chapeau de Paille (migmatites du complexe de base), le groupe de Mékinac (groupe métasédimentaire de plate-forme et, vers l'est, de marge continentale), le groupe de Montauban (volcanogénique, marin) et le complexe de La Bostonnais (comprenant diverses roches basiques et ultrabasiques).

Deux isogrades de métamorphisme ont pu être tracés. L'apparition de l'hypersthène est accompagnée d'une transformation profonde de la roche qui prend une teinte verte et les caractères bien connus des roches charnockitiques. Le "front de charnockitisation" est indépendant des limites stratigraphiques et d'un métamorphisme antérieur et apparaît comme l'expression de dômes de chaleur en liaison avec un évènement tectonique affectant des roches déjà métamorphiques et donc pauvres en eau à des températures élevées et des pressions modérées.

L'apparition de la muscovite dans les gneiss à grain fin du groupe de Montauban est accompagnée vers le sommet du groupe de celle de gneiss nodulaires à quartz et fibrolite qui pourrait représenter la destruction d'un équilibre précédent à staurotide ou à disthène. Dans cette zone, les structures primaires sont encore bien conservées: métabasalte à coussinets, granoclassements des pyroclastiques, structures primaires de certains intrusifs du complexe de La Bostonnais, etc.

Bien que très faillé, le complexe de La Bostonnais a en partie les caractères d'un ensemble ophiolitique et représenterait des vestiges de croûte océanique dans une ligne de suture séparant deux continents antérieurement indépendants. L'un à l'ouest est bien connu par les travaux effectués dans les Adirondacks, en Ontario et dans l'ouest du Québec. L'autre, à l'est, "Quebecia", est moins bien connu.

Abstract

Four new groups are recognized in the Saint-Maurice area; the are: the Chapeau de Paille Complex (migmatites of the basal complex); the Mékinac Group (a metasedimentary group of the platform and, towards the east, of the continental margin); the Montauban Group (volcanogenic, marine); and the La Bostonnais Complex (comprising various basic and ultrabasic rocks).

Two metamorphic isograds can be traced. The appearance of hyperthene is accompanied by an intense transformation of the rock which takes on a greenish tint and the typical characteristics of charnockitic rocks. The "charnockitization front" is independent of stratigraphic boundaries and of a previous metamorphism it appears as the expression of heat domes linked with a tectonic event that affected some already metamorphic rocks (consequently poor in H_2O) and brought high temperatures at moderate pressures.

The appearance of muscovite in the fine-grained gneisses of the Montauban Group is accompanied, towards the top of the group, by that of nodular quartz-fibrolite gneiss which could represent the destruction of a previous equilibrium with staurolite or kyanite. In this zone, primary structures are still well preserved: pillows in metabasalts; graded bedding in pyroclastic rocks; primary structures in certain intrusives of the La Bostonnais Complex, etc.

Although much faulted, the La Bostonnais Complex has in part, the characteristics of an ophiolitic assemblage and would represent vestiges of oceanic crust in a suture between two formerly independent continents. The western one is well known through studies carried out in the Adirondacks, in Ontario and in western Québec. The other, to the east, "Quebecia", is not as well known.

INTRODUCTION

La région du Saint-Maurice, d'une superficie de 1 800 km^2 est située à l'intérieur des latitudes 46°45' et 48°00' et des longitudes 72°30' et 74°00' soit près du centre de la province géologique de Grenville.

Cette dernière, partie majeure d'une chaîne dont on trouve des vestiges du Mexique au nord de l'Europe, d'abord considérée comme très ancienne à cause de son métamorphisme élevé et de la complexité des plissements qui l'affectent et bien que possédant des roches archéennes, est datée au potassium – argon de 1 100 à 900 Ma. Dietz et Holden (1966), examinant les dépôts de marge continentale, la comparent à la chaînes des Appalaches.

Wynne-Edwards qui a cartographié une grande région juste à l'ouest de la nôtre (Wynne-Edwards et al., 1966), établit un schéma d'évolution s'échelonnant de l'Archéen au Phanérozoïque (Wynne-Edwards, 1969). Selon lui un seul long épisode tectonique et métamorphique a affecté le soubassement cristallin ancien et sa couverture. Il subdivise cependant le Grenville en segments à caractères différents (Wynne-Edwards, 1972), la région du Saint-Maurice se trouvant dans le "Central Granulite Terrain". En résumé il envisage, 1 – à partir de 1 700 Ma, un dépôt de plus de 8 000 m d'épaisseur de sédiments et volcaniques (supergroupe de Grenville) sur un soubassement archéen ou aphébien, 2 – vers 1 400 Ma, la mise en place de grandes masses stratiformes d'anorthosite et de mangérite au voisinage du contact,

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3 – métamorphisme et déformation culminant vers 1 200 Ma, 4 – érosion et soulèvement très importants 15 à 20 km découvrant des zones où sont imprimées les déformations kénoréenne (pendage moyen 50° vers le N5°E), hudsonienne (direction NNW, pendage subvertical), et grenvillienne (pendage SE).

D'un autre côté, le rôle de la tectonique des plaques, bien étudié pour les périodes récentes a été reconnu, avec des modalités quelques peu différentes, au Précambrien (Dewey et Spall, 1975; Dewey, 1976; Haigraves, 1976). Burke et Dewey (1973), s'appuyant en partie sur une comparaison avec la surface lunaire, proposent, 1 – une première phase très mobile au début de l'Archéen faisant disparaître la plupart des premières roches différentiées, 2 – suivie d'une phase de transition de 2,6 à 1,8 milliards d'années qui serait propice à la formation d'aulacogène, 3 – tandis que la période suivante voit se développer la tectonique des plaques telle qu'elle est encore active actuellement, avec localement des collisions continentales entraînant un chevauchement des continents et la formation d'une ligne de suture complexe et généralement peu visible (Dewey, 1977).

Baer (1977), se basant sur des études pétrologiques, pense qu'à cause de l'instabilité de l'éclogite aux degrés géothermiques élevés du début du Précambrien, les zones de subduction n'ont pu exister avant un milliard d'années. Ce phénomène peut-être valable pour l'Archéen où les roches volcaniques basiques, marines et les pyroclastiques sont abondantes et correspondent d'après Burke et al. (1977) à un plus grand nombre de points de dissipation de chaleur. Cependant, au Québec, le rétrécissement par compression de plus de la moitié de cette croûte archéenne vers 2 500 Ma pour former le noyau du bouclier Canadien, et l'aspect différent avec sédimentation plus abondante du Protérozoïque, indique que le cycle de Wilson a dû débuter à cette époque; présence d'ophiolite (Dimroth, 1971; Leblanc, 1976) et de suture (Burke et al., 1977).

D'après Burke et al. (1976), l'existence de ligne de suture dans la province de Grenville est confirmée par la jonction des traces des pôles paléomagnétiques. Ils notent cependant. le manque d'évidence géologique d'une ligne de suture témoin d'un mouvement relatif entre un continent "Grenvillia" et le reste de l'Amérique du Nord "Laurentia", hypothèse proposée par Irving et al. (1972) et Palmer et Carmichael (1973). En effet, d'après Buchan et Dunlop (1976), les paléopôles de la partie sud du Grenville lui sont spécifiques tandis qu'au nord, des unités préorogéniques sont identifiées jusqu'à 100 km à l'intérieur de la province de Grenville. Un tracé préliminaire d'une ligne de suture est proposé par Irving et al. (1974), tandis que Thomas et Tanner (1975) et Brown et al. (1975), donnent des précisions sur ses extrémités est et ouest. Ueno et Irving (1976) considèrent qu'un soulèvement de la partie nord du Grenvillia vers 1 000 Ma est responsable d'une magnétisation particulière. Ce soulèvement, conséquence d'une collision continentale de type Tibétain (Dewey et Burke, 1973), aurait affecté, d'après Baer (1976b) des roches de remplissage d'aulacogène (groupe de Grenville, à l'ouest de notre région).

Pour étudier l'histoire métamorphique de la région du Saint-Maurice, il est nécessaire d'avoir des précisions sur l'origine des roches étudiées, leur appartenance ou leur relation avec des types connus. Pour celà, il est nécessaire de savoir jusqu'à quel degré les roches peuvent être modifiées au cours de leur histoire. Pour de La Roche (1974), la survivance des caractères géochimiques des ensembles métamorphiques témoigne de leur histoire antémétamorphique, même dans la catazone (de La Roche et al., 1971).

Le but de cette note est, après avoir mentionné les unités géologiques voisines déjà connues, de décrire les groupes et complexes spécifiques à la région du Saint-Maurice, de rassembler les critères sur leurs origines en s'appuyant sur la survivance de leurs caractères primaires géochimiques (analyses) et structuraux (basalte à coussinets) et de donner un aperçu de leurs histoires par leurs métamorphismes.

Cadre géologique

La région du Saint-Maurice est presque entièrement formée de gneiss et de migmatites, par contre les ensembles géologiques qui l'entourent (fig. 1) occupant des surfaces assez semblables, comprennent une certaines proportion d'intrusifs acides ou basiques. Ce sont: au sud-ouest, le groupe de Grenville; au nord, le groupe de Saint-Félicien et l'anorthosite du lac Saint-Jean; à l'est, le complexe du Parc des Laurentides.

Le groupe de Grenville est le plus anciennement connu. Logan (1863) le décrit comme un ensemble constitué principalement de calcaire cristallin et de quartzite reposant sur des gneiss monotones attribués à l'Archéen. Engel (1956) donne plus de précision sur ce groupe, auquel il ajoute une certaine proportion de roches volcaniques et de conglomérat. Wynne-Edwards (1972) lui accorde une épaisseur de 8 à 10 km. Dans les Adirondacks, McLelland (1974) estime à 6 km l'épaisseur des sédiments de plate-forme, distribués en quatre formations qu'il rattache au groupe de Grenville. Associés à ces métasédiments on trouve des massifs d'anorthosite et de roches intrusives charnockitiques et granitiques. Un de ces complexes, non loin de l'extrémité sud-ouest de notre région, est considéré par Schrijver (1975) comme la racine déformée d'un diapir composé. Des roches du groupe de Grenville affleurent dans la partie sud-ouest de notre région. Elles sont caractérisées par des compositions nettement métasédimentaires.

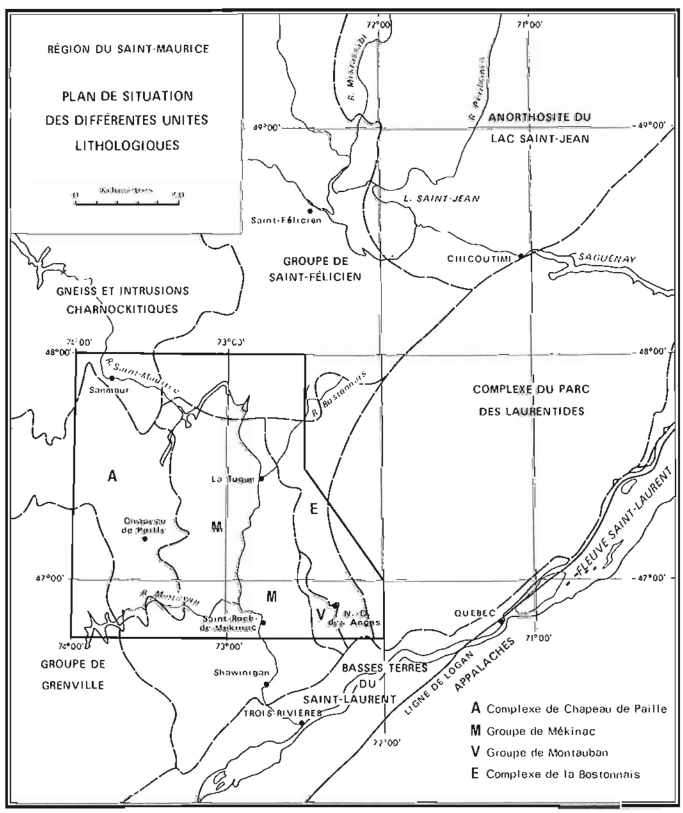

Figure 1. Plan de situation des différentes unités lithologiques de la partie centre-sud de la province de Grenville.

Le groupe de Saint-Félicien, mentionné pour la première fois par Engel (1956), désigne un ensemble de roches comprenant peu de calcaire cristallin, des quartzites impurs et une abondance de gneiss feldspathique et d'amphibolite visibles dans la région de Saint-Félicien sur le lac Saint-Jean. Une partie de ces gneiss a été cartographiée par Benoit (Benoit et Valiquette, 1971) directement au nord de notre région. Ce groupe apparait aussi au contact de l'anorthosite du lac Saint-Jean (Berrangé, 1962; Rondot, 1963). Parmi les roches intrusives citons: la grande masse d'anorthosite du lac Saint-Jean à laquelle il semble qu'il faille rattacher la plupart des petites masses anorthositiques affleurant au nord de notre région.

Dans notre région, ces mêmes gneiss de composition principalement granitique et granodioritique et à grain fin à moyen, apparaissent jusque dans le nord de la région de la Croche (Newham, 1973; Rondot, 1964), mais aux intrusions charnockitiques et granitiques, il faut ajouter des petites masses de péridotite, gabbro et diorite.

Le complexe du Parc des Laurentides, à l'est de notre région, est formé principalement de roches intrusives charnockitiques (Rondot et Marleau, 1977). S'étendant de Rivière-à-Pierre jusqu'au-delà du Saguenay, ce complexe ne contient que peu de gneiss et de métasédiments, mais plusieurs petits massifs anorthositiques et d'importants massifs charnockitiques, certains de ces massifs étant bien individualisés sur les cartes aéromagnétiques. Le granite de Rivière-à-Pierre, dans le coin sud-est de notre région, est l'un de ces massifs.

Travaux effectués

La région du Saint-Maurice (fig. 1 et 4) a été cartographiée à partir de 1953 (Tiphane, 1954; Klugman, 1956, 1963; Newham, 1973; Rondot, 1962, 1964, 1967, carte et texte en préparation). En 1976, le travail de reconnaissance stratigraphique et de synthèse avait pour but de présenter une carte à petite échelle (1/100 000), d'essayer de grouper les différentes formations de la région et d'en établir la stratigraphie.

Une première échelle stratigraphique est esquissée lors de la cartographie de la région de Mattawin-Lac-Chat-Est au 1/50 000 (Rondot, 1962, p. 101). Dans ce texte quatre niveaux métasédimentaires A, B, C, D se détachent d'un ensemble de migmatite monotone homogène ou rubannée (avec lits ou lentilles d'amphibolite-pyribolite) chaque niveau étant caractéristique d'un type de sédimentation: A – sédiments détritiques grossiers; quartzite, leptynite, conglomérat?, B – sédiments détritiques fins et chimiques; gneiss quartzeux, roches calcosilicatées, C – sédiments argileux et variés; gneiss à sillimanite, D – sédiments clastiques fins; gneiss à grenat, muscovite et amphibolite. Au sud-ouest de cette région, Philpotts (1967, p. 15) a retrouvé indépendamment les trois premiers niveaux, le niveau C servant de transition entre ses "Grenville inférieur" et "supérieur", et à l'est de cette région, Pyke (1967) reconnaît, au milieu du niveau D, des basaltes à coussinets (fig. 2). Karen Stamatelopoulou-Seymour (1975) étend l'origine volcanique à une partie des gneiss à grain fin et au minerai de Montauban.

Le travail de 1976 sur le terrain a permis d'avoir une idée un peu plus précise de ces niveaux, rassemblés en deux nouveaux groupes; Mékinac (A, B, C) et Montauban (D), de les suivre plus au nord, de reconnaître leurs limites et les unités qui les entourent et en particulier, à l'ouest, l'unité sur laquelle ils reposent; complexe de Chapeau de Paille, et, à l'est, un complexe basique assez particulier; complexe de La Bostonnais. Pour ce faire, outre la reconnaissance sur le terrain, on a utilisé la photogéologie, en particulier dans la partie est de la région. Chaque paire stéréoscopique comprenait, d'un côté les pendages figurés au lieu de leur observation (fig. 3). De l'autre, étaient tracés d'abord les failles, subrectilignes et, en général, discordantes avec les pendanges, puis les lignes structurales et contacts géologiques qui sont ensuite reportés sur les minutes pour vérification et correction. Le résultat a été la suggestion de nouveaux groupes (Rondot, 1976a) qui ont pu être décrits (Rondot, 1976b) et examinés par les membres du "club Grenville" lors de leur excursion annuelle et qui sont proposés ici à l'occasion de leur rapport avec le métamorphisme de la région.

DESCRIPTION DES NOUVEAUX GROUPES

Pour se conformer aux règles de l' "International subcommission on stratigraphic classification" (1972) chaque groupe sera décrit d'après les affleurements de sa localité type à l'échelle de ce travail d'exploration. Une étude plus détaillée permettrait d'isoler des formations et de montrer des variations latérales de faciès, cependant à ce stade de la reconnaissance, nous décrirons les groupes en fonction des niveaux repères et de l'origine qu'ils suggèrent.

Figure 2.

Coussinets du métabasalte de l'unité V_b. Photographie de D.R. Pyke. (GSC 203364)

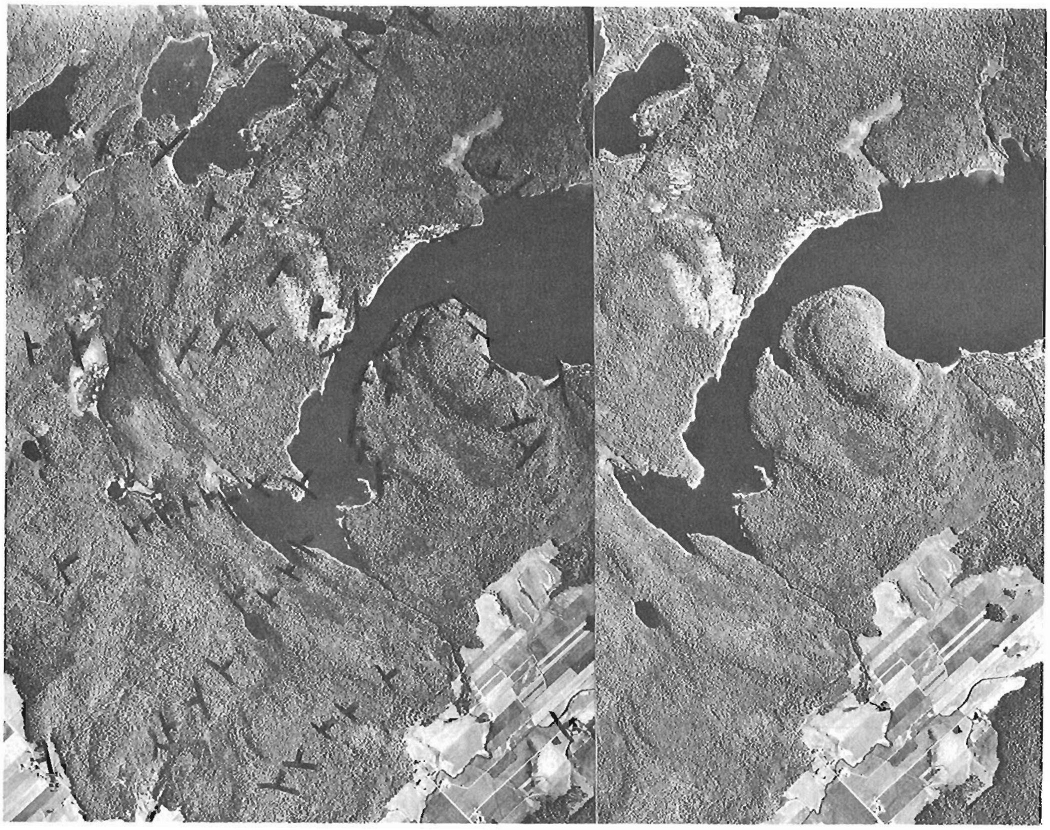

Figure 3.

Paire stéréoscopique avec pendages. Photographies au 1/40 000 du ministère des Terres et Forêts Q66308 – 6 et 7. Lac Blanc à 5 km à l'est de Montauban. (GSC 203364-A)

A- *Complexe de Chapeau de Paille* (nouveau)

Le terme de complexe de Chapeau de Paille désigne l'ensemble des roches qui affleurent dans la partie ouest de la région (fig. 4) entre la rivière Mattawin et Sanmaur. Le lieu-dit "Chapeau de Paille" est situé au nord du lac du même nom dans le centre ouest de la région (47°11'N, 73°32'W). Le nom apparaît même sur les cartes au 1/1 000 000, quelquefois sous le terme de "dépôt du Chapeau de Paille". Certaines cartes indiquent cependant un tracé de route qui n'est plus valable, l'accès se faisant maintenant à partir de Mattawin.

Les deux unités les plus importantes de ce complexe sont des migmatites grises et roses, à biotite-hornblende, qui affleurent le long de la route de Chapeau de Paille et d'autres chemins secondaires de la région. Bien que ces migmatites forment près de 90% des roches du complexe, on observe aussi, 1 – de minces niveaux, quelques centimètres, à quelques mètres, de quartzites impurs et de roches carbonatées et des masses irrégulières et peu importantes, 2 – de gneiss à grenat, gneiss à hornblende, amphibolite et métagabbro et 3 – de gneiss oeillé et de granite porphyroïde. Font aussi partie de ce complexe, des roches de même composition chimique mais à feldspath vert et pyroxène dans le centre de la région (fig. 4).

Ces roches se prolongent à l'est et sont semblables à celles du "complexe de base" de Wynne-Edwards (1966), en particulier ses unités, 1 – gneiss granitique et 1a – gneiss à oligoclase, et aux roches décrites par Osborne (1935) sous le nom de "Lacoste Series".

Les deux unités principales du complexe semblent aussi abondantes l'une que l'autre. Cependant, là où elles affleurent l'accès est difficile et il n'existe que des chemins temporaires de compagnie de bois. La localité type (environ du lac du Chapeau de Paille) a été choisie à cause de sa proximité de Mattawin, principal accès à cette région. Les migmatites grises, A_1, affleurent près des anciennes constructions (47°11'N, 73°32'W) tandis que les migmatites roses, A_2, sont abondantes le long du chemin à 3 km au nord-ouest de cet endroit (47°12'N, 73°34'W). Cet établissement a été très actif autrefois et il y avait même une piste d'atterrissage, mais les constructions sont actuellement abandonnées et, au cas où le chemin disparaîtrait lui aussi, les affleurements d'un autre chemin, se rendant de Mattawin au village indien de Manouane peuvent servir de site de référence. Les migmatites roses, A_2, affleurent juste à l'ouest du pont de la rivière à la Chienne au nord du lac du même nom (47°03'N, 73°31'W) et les migmatites grises, A_1, à l'est du lac du Chevalier (47°02'N, 73°42'W). Tous ces sites sont sur la feuille au 1/50 000 n° 31P4 Est, du ministère de l'Energie des Mines et des Ressources d'Ottawa.

Les migmatites grises, A_1, sont des roches gneissiques, à grains moyen à grossier, hétérogènes à l'échelle de l'échantillon du fait d'une répartition en amas, lits ou lentilles des minéraux mafiques, et la présence occasionnelle de filonnets de quartz et feldspath rose, mais de composition monotone à l'échelle de l'affleurement ou de la région. La proportion de quartz varie de 10 à 40%. Le microcline peut manquer, mais il représente habituellement de 5 à 15% du volume de la roche. L'oligoclase calcique, An 18 à An 31, peu maclé, est le minéral le plus abondant. La hornblende verte apparaît dans presque tous les échantillons tandis que la biotite brune à brun-noir est constante, ces deux minéraux représentant plus de 10% du volume de la roche. Les autres constituants sont les minéraux opaques, les minéraux accessoires; apatite, zircon et sphène (rare) et les minéraux d'altération; chlorite, épidote, alanite, séricite.

Les migmatites roses, A_2, sont des roches gneissiques à grain généralement grossier. La roche est hétérogène à l'affleurement et on note des passages plus riches en minéraux mafiques ou en gneiss quartzeux ou de composition variée. La roche typique contient de 20 à 40% de quartz, 25 à 50% de microcline, 15 à 40% d'oligoclase sodique An 10 à

An 20, 0 à 5% de hornblende vert sombre, 1 à 3% de biotite brun-vert et 1% de minéraux opaques. Les minéraux accessoires sont variés; apatite, zircon, sphène, allanite, tourmaline. Il y a aussi occasionnellement des minéraux d'altération et du grenat.

Les compositions moyennes de ces deux types de roche sont les suivantes:

Type et nombre d'échantillons

	A_1 (13 éch.)	A_2 (12 éch.)
Quartz	24%	26%
Microcline	9%	39%
Oligoclase (An 18 à 31)	54%	(An 10 à 20) 29%
Hornblende	4%	2%
Biotite	7%	2%
Accessoires et opaques	2%	2%

En général, ces deux types de roches occupent des zones distinctes, cependant, à cause des plissements très complexes il n'est pas possible actuellement de connaître leur répartition et de mesurer leurs épaisseurs. On note, d'après la carte, que les zones de migmatites roses sont à l'intérieur des zones de migmatites grises, mais ce fait peut-être dû à la transformation du microcline rose en microperthite blanche à l'approche des roches charnockitiques.

Du point de vue stratigraphique, on ne trouve aucune roche en dessous des roches du complexe qui sont dans le prolongement des roches que Wynne-Edwards (1969) considère comme archéennes. Les roches qui sont stratigraphiquement au-dessus, si on se réfère aux pendages les plus fréquents, sont, au sud-ouest, les roches du groupe de Grenville qui débutent par des gneiss à sillimanite et des gneiss rouillés, se poursuivent presque sans interruption par des roches possèdant des composition spécifiques des roches sédimentaires surtout quartz et carbonates. Cependant, on ne peut dire avec précision s'il y a discordance entre ces deux groupes. Au nord, le complexe disparaît dans les roches charnockitiques gneissiques ou intrusives. A l'est, toujours si on se fie au pendage le plus fréquent, le complexe passe sous les roches du groupe de Mékinac en particulier sous les épaisses strates de métagabbro à grenat de la base du groupe, sans que l'on puisse dire, là encore, qu'il s'agit d'une discordance stratigraphique.

L'origine des roches de ce complexe, du fait de leur composition monotone, est assez difficile à préciser.

M- Groupe de Mékinac (nouveau)

Le groupe de Mékinac rassemble les trois premiers niveaux métasédimentaires reconnus dans la région de Mattawin-Lac-Chat-Est (Rondot, 1962) et les roches dans lesquelles on les trouve soit, à la base des strates assez étendues de métagabbro à grenat et de grandes épaisseurs de migmatite à biotite, hornblende ou à pyroxènes. Le terme de Mékinac désigne à la fois un canton, un lac, une rivière et deux villages (Saint-Roch et Saint-Joseph) tous situés sur les roches de ce groupe, à une trentaine de kilomètres au nord des villes de Shawinigan et Grand-Mère sur le Saint-Maurice.

En 1958 et 1959 dans la région de Mattawin-Lac-Chat-Est, les niveaux A, B et C ont été reconnus avec suffisamment de certitude pour établir une première colonne stratigraphique. L'erreur sur les épaisseurs mesurées, en particulier dans les migmatites charnockitiques est cependant assez grande. Lors de la revision en 1976, on a essayé de suivre avec plus de précision les différents niveaux stratigraphiques grâce à l'interprétation photogéologique et à les rechercher dans les régions avoisinantes déjà cartographiées. Ce qui a amené à présenter la colonne stratigraphique de la figure 5. Le groupe de Mékinac y est subdivisé en quatre unités séparées par les niveaux A, B et C. Pour les descriptions de plusieurs sites où apparaissent ces niveaux on peut se reporter au rapport géologique de cette région (Rondot, 1962 DPV 373). On décrira, cependant ici, les coupes-types à des endroits aisément accessibles et, sauf pour le niveau B (falaise à saphirine du Saint-Maurice), différents de ceux décrits dans le DPV 373.

Unité M_1

L'unité inférieure est caractérisée par la présence d'épaises couches de métagabbro à grenat. Certains ont gardé leur plagioclase basique et des restes de structures ophitique. La roche est en générale massive et à grain fin à moyen, cependant, localement elle est foliée et hétérogène et elle passe à une anorthosite gabbroïque à feldspath laiteux. Ces métagabbros se présentent en strates ayant jusqu'à 100 m d'épaisseur et on peut les suivre sur plus de 10 km lorsque la roche, massive et homogène, reste en relief sous forme de collines basses et allongées.

Les compositions minéralogiques en volume pour cent sont les suivantes:

Compositions minéralogiques des métagabbros

	(1)	(2)	(3)
Plagioclase	(An 50 à 55) 89	(An 38 à 66) 42	(An 38 à 58) 28
Hypersthène	-	0 à 15 moy. 7	9
Augite	2	0 à 30 moy. 6	14
Hornblende	4 verte	32 brun-vert	35 brune
Biotite	-	0 à 6 moy. 1	traces
Grenat	3	9	0
Opaques	1	3	5
Accessoires	1	peu	peu
Total de ferromagnésiens	10%	58%	72%

(1) métaanorthosite gabbroïque moyenne de 3 estimations

(2) métagabbro moyenne de 10 estimations

(3) métagabbro moyenne de 4 estimations

La localité type décrite dans le livret-guide de l'excursion du "club Grenville" (Rondot, 1976b) est située à l'ouest du Saint-Maurice sur le chemin qui, du bac de Mattawin, donne accès à cette partie de la région (Chapeau de Paille, Manouane etc.). L'affleurement est à 3 km (2,2 milles) au sud-ouest de la barrière du Parc du Saint-Maurice, soit à 13 km (7,9 milles) du lac, sur la feuille au 1/50 000 – 31 I 14 Est, là où le chemin accède à une petite colline (73°31'W, 46°54'N). L'affleurement est continu sur une assez grande distance mais ne permet pas de mesurer directement l'épaisseur de la strate. Sur cet affleurement on peut observer plusieurs structures; sub-ophitique, foliée, massive etc., et une certaine hétérogénéité dans la composition minéralogique.

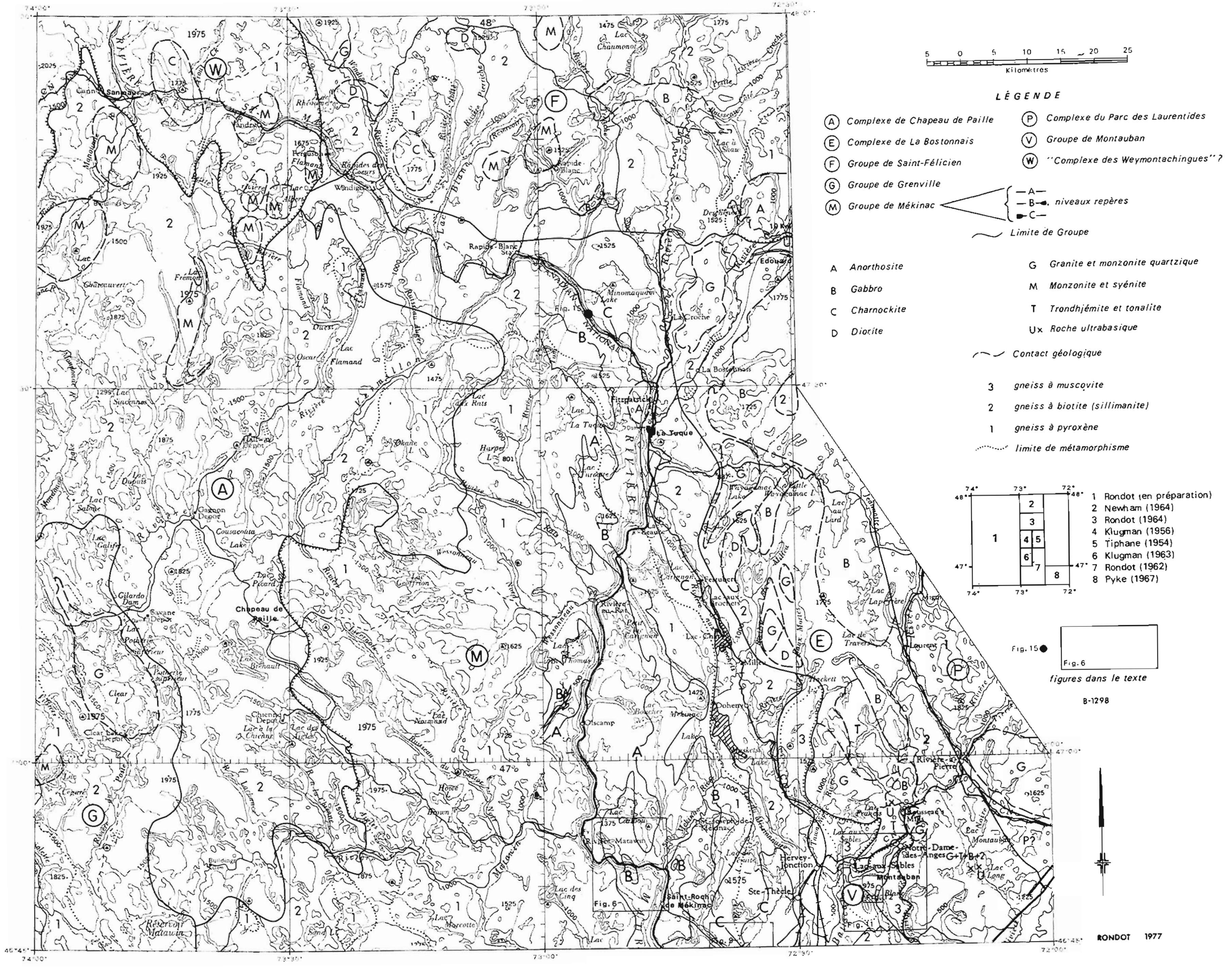
Kilomètres
LÉGENDE
A Complexe de Chapeau de Paille
E Complexe de La Bostonnais
F Groupe de Saint-Félicien
G Groupe de Grenville
M Groupe de Mékinac
P Complexe du Parc des Laurentides
V Groupe de Montauban
W "Complexe des Weymontachingues"?
—A— —B—. —C— niveaux repères
Limite de Groupe
A Anorthosite
B Gabbro
C Charnockite
D Diorite
G Granite et monzonite quartzique
M Monzonite et syénite
T Trondhjémite et tonalite
Ux Roche ultrabasique
Contact géologique
3 gneiss à muscovite
2 gneiss à biotite (sillimanite)
1 gneiss à pyroxène
limite de métamorphisme
1 Rondot (en préparation)
2 Newham (1964)
3 Rondot (1964)
4 Klugman (1956)
5 Tiphane (1954)
6 Klugman (1963)
7 Rondot (1962)
8 Pyke (1967)
Fig. 15 Fig. 6
figures dans le texte
B-1298
RONDOT 1977
La Tuque
Chapeau de Paille
Rapide-Blanc
Saint-Roch de Mékinac
Notre-Dame-des-Anges
Montauban
Lac aux Sables
Hervey-Jonction
Ste-Thècle
Réservoir Mataw̄in

Figure 4. (page précédente)

Carte des groupes et complexes et des zones de métamorphisme de la région du Saint-Maurice.

La route de La Tuque qui longe le Saint-Maurice de part et d'autre de Mattawin (feuille 31 I 15 Ouest) recoupe les différents types de roches de l'unité M_1 et, bien que le métagabbro n'affleure pas immédiatement au bord de la route, on peut l'observer ainsi que l'anorthosite gabbroïque sur une petite colline (72°54'W, 46°52'N, n° 1 de la figure 6) à 4 km au sud de Mattawin qui peut servir de site de référence.

Les autres types de roche de l'unité sont des amphibolites et pyribolites en couches ou lentilles d'épaisseurs irrégulières dans les migmatites charnockitiques, qui représente 80 à 90% des roches du voisinage, que l'on peut observer le long de la route entre le village de Mattawin et le bac (72°55'W, 46°54'N). Les compositions moyennes sont les suivantes pour (1) les migmatites charnockitiques vert pâle (moyenne de 10 estimations) (2) les migmatites charnockitiques hétérogènes (moyenne de 15 estimations et (3) une enclave de pyribolite:

Le niveau A (cf. DPV 373 p. 69 et suivantes) comprend surtout des roches quartzeuses et des leptynites (en partie à grenat) et des gneiss calciques (scapolite). A la base ce sont surtout des roches quartzofeldspathiques et des gneiss très quartzeux à grain grossier qui dominent. Au nord du lac Attraction, on note même un affleurement d'aspect conglomératique à plus de 50% de quartz, l'autre minéral étant le feldspath potassique. Plus haut alternent amphibolite et gneiss quartzeux en lits minces, par endroits bien stratifiés et de composition non uniforme. Ceux-ci passent progressivement aux migmatites charnockitiques.

L'accès à cette bande n'est pas aisé là; où elle traverse la route et la rivière il n'y a pas d'affleurement. Aussi faut-il chercher la coupe-type à deux kilomètres de la route en empruntant un chemin privé (Robert Mongrain) qui passe près du lac Daniel (feuille de Mattawin 31 I 15 Ouest, 72°53'W, 46°53'N) le long d'un escarpement à moins d'un kilomètre à l'est du lac (n° 2 de la figure 6). Le pendage étant oblique par rapport à l'allongement de l'escarpement on peut observer une coupe continue de la base (au nord-ouest) au sommet du niveau (terminaison sud-est de la falaise). Cependant l'aspect conglomératique de la base et la bonne stratification des niveaux supérieurs ne sont peut-être pas visibles ici.

Compositions minéralogiques des migmatites charnockitiques

	(1)	(2)	(3)
Quartz	22,5	22,4	-
Perthite	19,6	14,1	-
Plagioclase	(An 25 à 30) 48,6	(An 25 à 36) 42,6	(An 38) 40
Hypersthène	1,1	2	10
Augite	3,3	2,6	15
Hornblende	2,1	6,6	30
Biotite	0,5	4,4	-
Opaques	1,5	3,6	5
Apatite	0,5	1	peu
Autre	0,3	0,7	-
Total de ferromagnésiens %	8,8%	19,9%	60%

Outre ces roches on observe aussi dans cette unité quelques amas de carbonates et roches calcosilicatées, et des injections et des pegmatites alaskitiques.

La limite inférieure de l'unité est, arbitrairement, la base de la première strate de métagabbro à défaut de l'observation d'une discordance avec l'unité précédente. La limite supérieure est la première apparition d'un banc continu de métasédiments (niveau A). L'épaisseur d'un kilomètre attribuée à cette unité peut-être erronnée ou ne présenter qu'une moyenne d'épaisseur variable d'un point à un autre de la région. Elle est en effet très étendue allant de Mattawin au lac Flamand (à 50 km au nord-ouest de La Tuque). Aux extrémités de cette bande, les métagabbros sont cependant beaucoup plus rare ce qui pourrait indiquer que l'unité représente un ensemble volcanogénique local.

Unité M_2

L'unité suivante comprend, à la base, le niveau A (100 m ou plus) et une grande épaisseur (plus d'un kilomètre) de migmatite charnockitique avec deux ou trois niveaux de métagabbro avec ou sans grenat. Lors de la cartographie, en 1958, de la feuille de Mattawin, le niveau A a été repéré presque chaque fois qu'un itinéraire le traversait. Il représente en effet les seules roches franchement métasédimentaires de cette partie de la région.

Les migmatites charnockitiques qui forment l'essentiel de l'unité M_2 ne sont pas différentes de celles de l'unité inférieure. On y observe aussi des masses stratiformes de métagabbro (qui peuvent ne pas contenir de grenat) dont l'épaisseur ne dépasse pas 30 m. Les amphibolites-pyribolites sont par contre abondantes dans cette unité. Le long de la route allant de Mattawin à Saint-Roch de Mékinac, un affleurement particulier peut attirer l'attention. C'est une jotunite à gros plagioclases noirs et niveaux à magnétite tout à fait inhabituels (n° 3 de la figure 6). Il est difficile de savoir s'il s'agit d'un sill plus récent ou d'une roche contemporaine des métagabbros.

Il est aisé de tracer la limite inférieure de l'unité qui correspond à la première apparition des métasédiments. De même l'unité supérieure correspond à l'apparition d'une autre série métasédimentaire, épaisse. Ainsi, l'unité M_2 a pu être suivie de Saint-Roch de Mékinac jusqu'au nord-ouest de La Tuque avec une épaisseur à peu près constante.

Unité M_3

L'unité M_3 est plus riche en métasédiments typiques. La formation principale, le niveau B, en affleurement continu le long du Saint-Maurice, comprend plusieurs couches de quartzite, quartzite feldspathique ou grenatifère, gneiss quartzeux à carbonate, gneiss calcosilicaté, gneiss à diopside et roche à carbonate. Cette formation affleure aussi largement à l'est de Saint-Roch de Mékinac et jusqu'au lac Mékinac. On y observe cependant la présence de quartzites plus purs et du graphite dans les roches à carbonates et calcosilicates. La partie supérieure de l'unité contient aussi des métasédiments, en particulier de nombreux niveaux de roches calcosilicatées et quelques lentilles de quartzite.

La coupe type du niveau B est représentée par la falaise à saphirine du Saint-Maurice à 10 kilomètres en amont de Saint-Roch de Mékinac (n° 4 de la figure 6, 72°50'W, 46°52'N) elle a été décrite dans le DPV 373 et le livret guide de

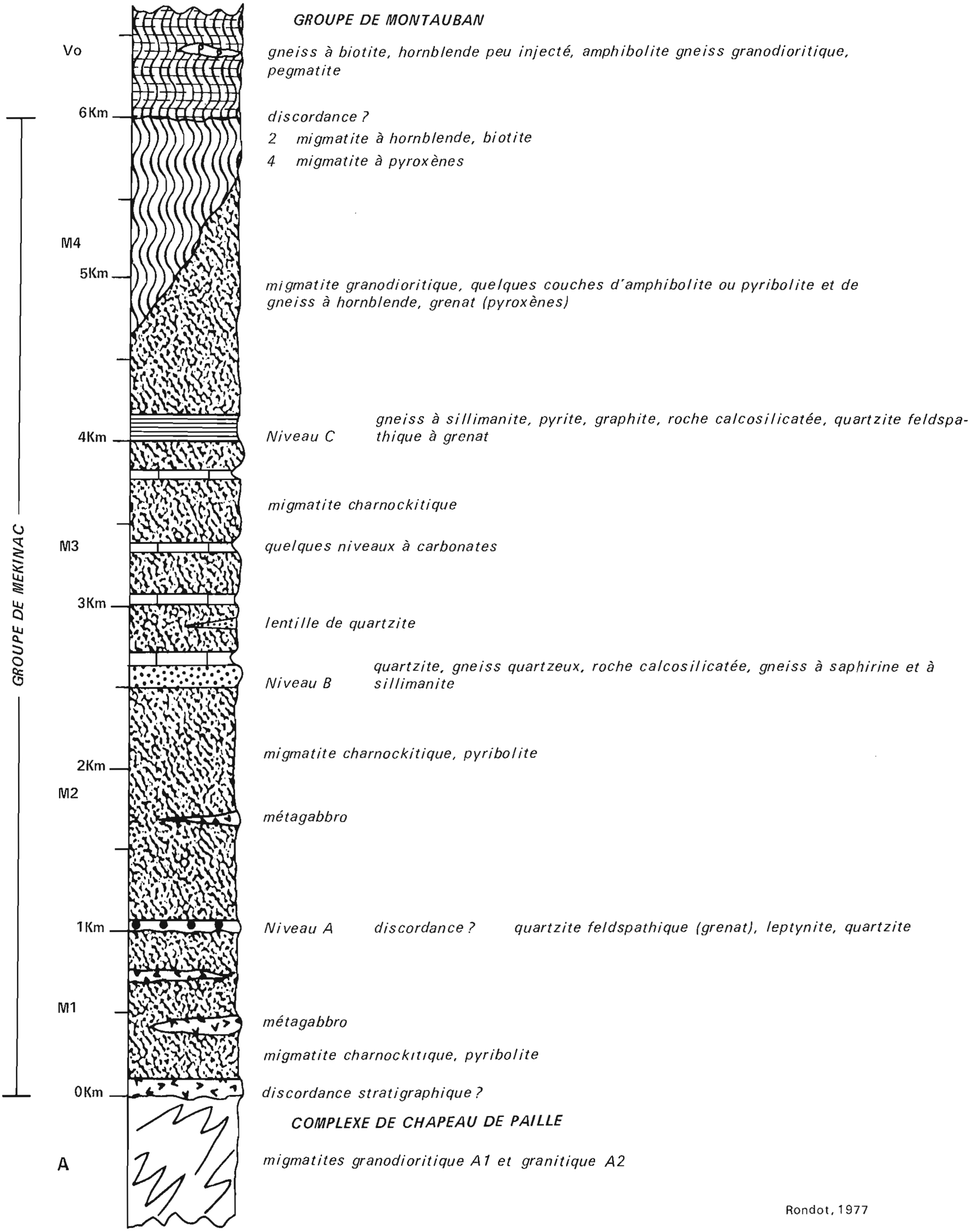

Figure 5. Echelle stratigraphique du groupe de Mékinac.

l'excursion du club Grenville en contient une coupe faite par Joël Brun (document disponible au ministère des Richesses naturelles de Québec) dont on trouvera un résumé en annexe 1. La coupe se trouvant près du coeur d'un synclinal (fig. 6) mais sur le flanc inverse, toutes les couches sont renversées et la base de la formation est à l'est de la coupe. A l'exception des niveaux carbonatés (fig. 7 parties claires) les différents types de roches sont bien stratifiés (fig. 8) et il n'y a pas de répétition tectonique. Il est à noter que cette formation composée de plus de 200 m de roches principalement quartzeuses et calcosilicatées avec intercalation de migmatites charnockitiques ne contient qu'un seul mince niveau de gneiss à sillimanite, où l'on trouve, aussi la saphirine (à l'extrémité gauche de la figure 7) étudiée par Ferguson (1974).

A 15 km au nord-nord-ouest de Mattawin, une couche de métacalcaire (calcite, dolomite) visible sur plus de 8 km reposant en synclinal sur un banc de quartzite de 2 à 3 m d'épaisseur pourrait appartenir au niveau B (fig. 4).

La partie supérieure de l'unité M_3, bien que principalement formée de migmatites charnockitiques, comprend plusieurs niveaux de roches calcosilicatées et quelques lentilles (quelques mètres d'épaisseur sur 10 à 20 m) de quartzite. Les migmatites charnockitiques sont localement assez riches en minéraux ferromagnésiens et plus pauvres en quartz. Elles passent à des amphibolites-pyribolites plus ou moins homogènes qui pourraient provenir d'anciens métasédiments.

La limite inférieure de l'unité est bien repérable sur le terrain par l'apparition brusque de couches de métasédiments quartzeux et calcosilicatés et la limite supérieure par celle de métasédiments alumineux (niveau C). L'origine des sédiments de cette unité indique une transgression avec, à la base des sédiments quartzeux et magnésiens (gneiss à diopside), plus haut des carbonates pour aboutir avec le niveau C, aux sédiments argileux. L'unité M_3 est parallèle et à peu près aussi étendue que l'unité M_2.

Figure 6. Fragment de la carte au 1/100 000 de la région du Saint-Maurice au nord de Saint-Roch de Mékinac 1, 2, 3, 4 sites décrits dans le texte, site 2– coupe-type du niveau A, 4– coupe-type du niveau B.

Unité M_4

L'unité supérieure du groupe de Mékinac comprend elle aussi deux parties. La partie inférieure nettement métasédimentaire est caractérisée par des couches de gneiss à sillimanite (graphite), la partie supérieure par d'épaisses migmatites granodioritiques assez monotones.

Les roches du niveau C affleurent largement sur plusieurs kilomètres carrés à l'est et au nord du lac Mékinac soit autour de la baie de la Croix (72°39'W, 47°02'N) et à l'ouest de Lac-Chat (72°39'W, 47°10'N) de la feuille 31 P 2 Est; deux endroits qui peuvent servir de sites de référence. Des descriptions pétrographiques de ces roches ont été données dans le DPV 373 (p. 82 à 90). Cependant les roches y sont très plissées, peut-être à cause de leur plasticité (gneiss alumineux) et faillées (proximité de failles régionales importantes).

La coupe-type du niveau C a été choisie dans la partie sud de la feuille de Mattawin 31 I 15 Est, là ou réapparaît ce niveau après l'interruption de grandes masses de diorite, parce que le pendage y est régulier sur le flanc ouest d'un synclinal renversé au sud du lac Archange (fig. 9, site 12, 72°38'W, 46°46'N) et que les couches ne semblent pas présenter de répétition tectonique. Cependant, à cause d'un paysage en dents de scie, les affleurements n'apparaissent guère que dans les cuestas. On trouvera, en annexe II, la succession de faciès estimée à partir de deux itinéraires. L'élément caractéristique de ce niveau est une couche de 30 m de gneiss à sillimanite bien stratifiée.

La partie supérieure de l'unité M_4 est bien représentée à l'est du lac Mékinac (fig. 4) par des migmatites de composition monotone localement rubannées ou seule varie la proportion relative des minéraux suivants: quartz 0 à 40%, microcline 0 à 50%, plagioclase 10 à 70%, biotite 0 à 20%, hornblende 0 à 60%, grenat, peu répandu, 0 à 20%. Ces migmatites sont de teinte grise à rose.

La limite inférieure de l'unité, premières couches de roches quartzeuses du niveau contenant des gneiss à sillimanite, a été reconnue jusqu'à La Tuque et au delà. La limite supérieure, dernière accumulation de roches d'une épaisse série monotone ne peut-être tracée que par l'apparition de roches différentes; complexe intrusif de la Bostonnais, volcaniclastiques du groupe de Montauban. Cette unité M_4 a la même extension nord-sud que les précédentes. Son origine, nettement sédimentaire et différenciée à la base pour le niveau C, avec peut-être intercalations volcaniques, devient

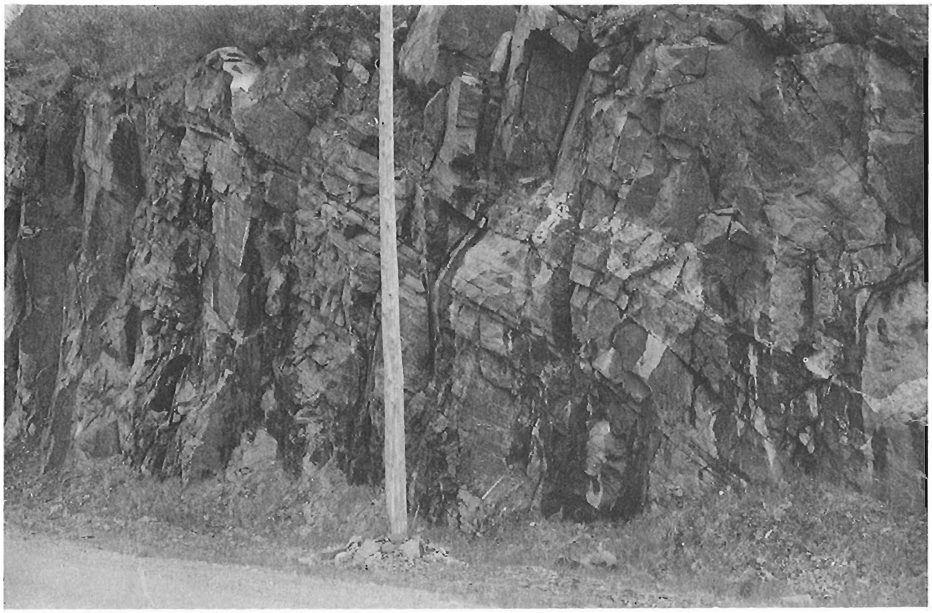

Figure 7.

Falaise à saphirine du Saint-Maurice. Niveau B, couche de calcaire cristallin (claire) dans les roches charnockitiques gneissiques (grises). (GSC 203364-B)

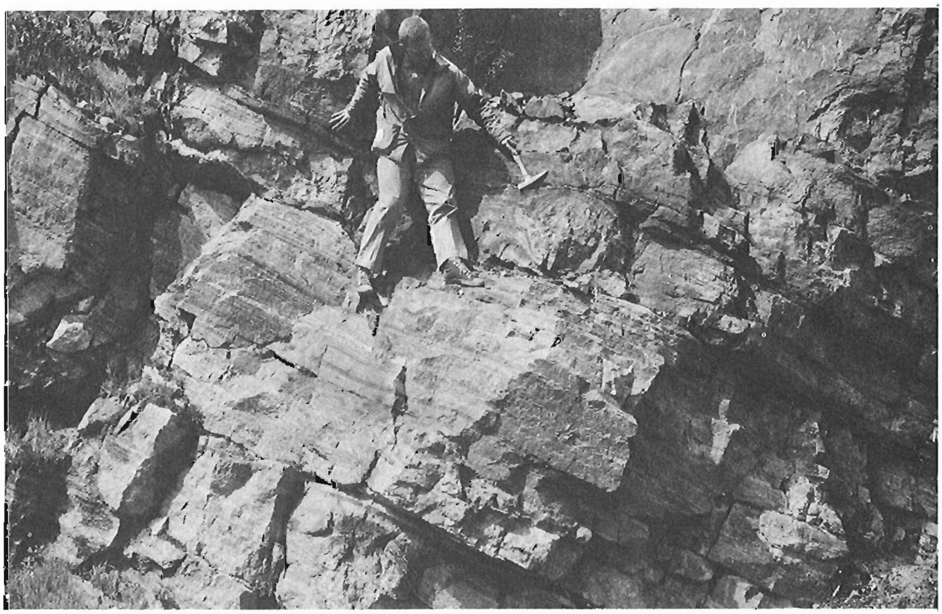

Figure 8.

Falaise du Saint-Maurice. Couches calcosilicatées succédant aux quartzites à grenat. Nodule à diopside et phlogopite. (GSC 203364-C)

immature vers le sommet avec de grosses accumulations (2 000 m?) de migmatites de composition rappelant celle de grauwacke.

Des dykes de composition intermédiaire à basique recoupent les roches de cette unité au nord de Saint-Tite.

V- *Groupe de Montauban* (nouveau)

Le groupe de Montauban (fig. 10) est caractérisé par des gneiss à grain fin, avec injections quartzofeldspathiques à la base seulement où l'on observe aussi d'épaisses pegmatites à grenat. Ces gneiss contiennent, pour la plupart, de la pyrite, ce qui leur donne une teinte rouillée à l'affleurement. Ils sont par endroits très bien lités surtout dans les unités supérieures du groupe où l'on observe un classement minéralogique et par grosseur de grain. D'autres types de roche sont aussi caractéristiques de ce groupe; ce sont des quartzites, des amphibolites-métabasaltes, localement à coussinets et des zones à carbonates avec amas de sulfures étudiées par Pyke (thèse de doctorat, McGill, 1967) et Karen Stamatelopoulou-Seymour, thèse de maîtrise, McGill, 1975). La figure 10 est basée en grande partie sur les travaux de Pyke et de Smith (1956), réinterprétés à l'aide de la photogéologie stéréoscopique avec pendages. La figure 11, nous montres que les différents types de roches forment un synclinal complexe presque entièrement entouré par le niveau de quartzite. La figure 11, basée sur la cartographie de détail de Smith, nous révèle l'emplacement et l'épaisseur moyenne des différents niveaux qui ont été mesurés sur le terrain (annexe III) pour donner la colonne stratigraphique de la figure 12.

La limite inférieure du groupe est théoriquement assez difficile à tracer au milieu de migmatites granodioritiques grises mais les deux types de roches, de composition assez semblable, ont cependant été séparés lors de la cartographie de cette région, peut-être à cause du caractère de roche homogène à grain fin (si l'on excepte les injections) de l'unité V_0 du groupe de Montauban comparé à celui de roche hétérogène et à grain moyen des migmatites (M_4) du groupe de Mékinac et à la présence de nombreuses failles accompagnant le passage d'un groupe à l'autre. La limite supérieure du groupe de Montauban est une lacune stratigraphique due à la fois à l'érosion et aux discordances tectoniques.

L'origine volcanique marine de la plupart des roches du groupe de Montauban a été démontrée par Pyke (1967) et Karen Stamatelopoulou-Seymour (1975). Quarante analyses de roches de compositions variées du groupe de Montauban, prises dans ces thèses et traitées par ordinateur selon la classification d'Irvine et Baragar (1971), (J. Bourne, communication personnelle) donnent: 55% de tholéite, 35% de roches calcoalcalines et 10% de roches alcalines, ou encore; 55% de basalte, 10% d'andésite, 25% de dacite et 10% de rhyolite. Cette composition mixte suggère la proximité du continent comme l'indiquerait la présence du niveau de quartzite, à moins que ce dernier ne représente un ancien niveau de chert. D'autre part, si l'on groupe les roches analysées d'après les indices de différentiation de Thornton & Tuttle (1960), on s'aperçoit que la presque totalité (une exception sur 40) est, soit au-dessous de 35, soit au-dessus de 65. Cette disposition bi-modale indiquerait, selon Martin & Piwinskii (1972), une mise en place anorogénique en période de tension (rift), ce qui n'est pas le cas pour un arc insulaire.

Les nombreux dykes de roches à grain fin, de compositions variées, qui recoupent les roches du groupe Mékinac au nord de Saint-Tite (Rondot 1962, p. 145 à 150) pourraient être reliés, sinon directement à l'éruption des roches volcaniques marines du groupe de Montauban, du moins à des manifestations volcaniques pénécontemporaines, reprises ensuite par une période orogénique et métamorphique.

Les roches du groupe de Montauban ne forment qu'une bande étroite (10 km) qui s'étend sur cinquante kilomètres, du lac Hackett au basses-terres du Saint-Laurent.

La Bostonnais est un petit village situé à 10 km au nord-est de La Tuque à proximité duquel affleurent les différents faciès du complexe (voir la carte du DPV 372 pour la distribution des différents types de roche et les cartes topographiques 31 P 10 Est et 31 P 9 Ouest pour la localisation des sites). A l'ouest et au nord du village, entre les rivières Croche et Bostonnais affleurent des roches gneissiques et massives autour et à l'intérieur d'un massif de monzonite quartzique, tandis qu'à l'est, les roches gneissiques et massives accompagnées de dykes de roches à grain fin bordent un grand massif de roches grenues basiques.

Le site le plus approprié pour observer les roches gneissiques et grenues, en petites masses, de ce complexe est à la sortie nord du village de La Croche (72°44', 3 W, 47°35', 5 N) où une petite falaise (fig. 13) nous permet d'examiner divers faciès gneissiques et massifs de compositions variées formant un mélange très complexe. La roche gneissique, en général riche en minéraux ferromagnésiens, a une composition trondhjémitique à dioritique. La roche massive, grise, à grain moyen à grossier est en général de composition dioritique.

Les roches du massif de monzonite quartzique affleurent au bord de la rivière Deschênes le long du chemin qui mène au lac du même nom (71°36', 5 W, 47°39', 5 N). Le long de la route et dans le massif même on note plusieurs textures allant de gneissique à porphyroïde. L'oligoclase sodique est en général plus abondant que le microcline.

Les roches basiques affleurent largement le long du chemin qui, de La Bostonnais (à 20 km au nord du village), mène au lac Edouard le long de la latitude 47°40'N. Après avoir passé une mylonite de la faille qui longe la rivière Bostonnais, on observe de nombreux affleurements de roches basiques à intermédiaires, homogènes et massives ou au

E- Complexe de La Bostonnais (nouveau)

La transition est brutale, entre les gneiss rouillés à grain fin à muscovite du groupe de Montauban et les roches qui affleurent plus à l'est. Il n'y a pas continuité stratigraphique et bien que des contacts d'intrusions soient visibles par endroits, il y a aussi de nombreux contacts par faille. Les roches qui affleurent à l'est de Montauban sont les même que celles que l'on observe plus au nord et jusqu'au delà de La Tuque. Elles comprennent une quantité à peu près égale de roches gneissiques et de roches massives de compositions allant d'acide à ultramafique et ayant conservées, en grande partie, leurs caractères pétrographiques primaires. Les roches basiques sont les plus abondantes dans ce complexe plus ou moins stratifié. Ces roches forment un ensemble que l'on peut délimiter et identifier assez facilement. Je les groupe sous le terme de complexe de La Bostonnais.

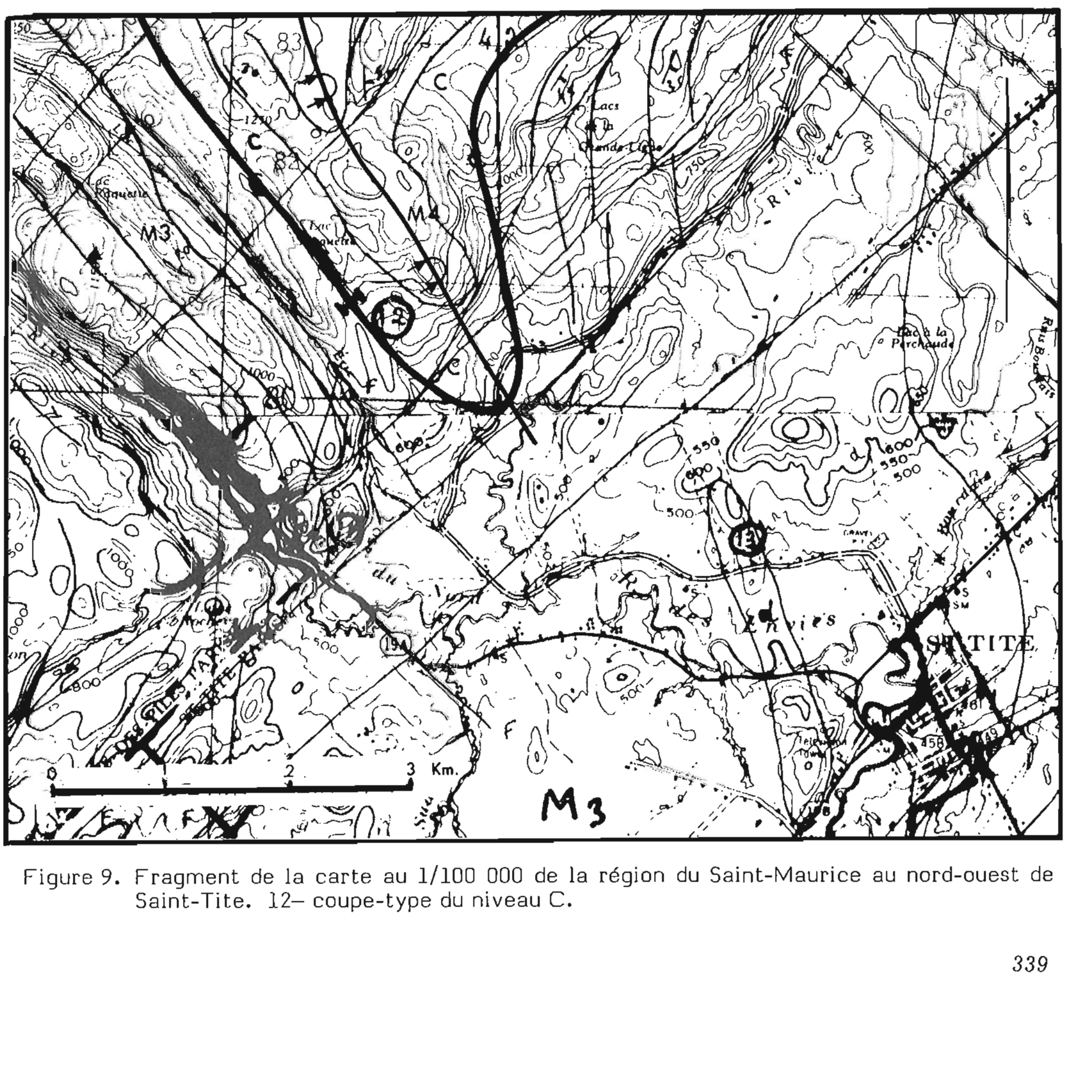

Figure 9. Fragment de la carte au 1/100 000 de la région du Saint-Maurice au nord-ouest de Saint-Tite. 12– coupe-type du niveau C.

Figure 10. Fragment de la carte au 1/100 000 de la région du Saint-Maurice autour de Montauban.

contraire hétérogènes et localement injectées de matériel alaskitique. Les différents faciès n'ont pas été suivis latéralement et on ne peut donc les situer les uns par rapport aux autres. A 10 km du pont sur La Bostonnais (72°23', 7 W, 47°39', 5 N) une masse de roche ultrabasique a été minée pour en extraire le nickel. On peut observer là plusieurs termes basiques à ultrabasiques: gabbro, pyroxénite à olivine, harzburgite, serpentinite, etc.

D'autres types de roches basiques de ce complexe affleurent le long des chemins forestiers qui joignent La Tuque à Rivière-à-Pierre. Il s'agit de gabbro et gabbro-anorthosite de teinte mauve. Le plagioclase, très maclé, est bourré de très fines inclusions.

Une dizaines de dykes de roche à grain fin de composition en général dioritique avec oligoclase calcique affleurent au nord et à l'est de La Bostonnais. Ces dykes sont peu épais (un mètre ou moins) et contiennent plus de 30% de minéraux ferromagnésiens, pyroxène, hornblende, biotite, opaques (DPV 372).

Les limites du complexe de La Bostonnais sont assez faciles à tracer puisqu'il s'agit pour plus de la moitié de roches intrusives généralement basiques et que les roches gneissiques qui les accompagnent sont à grain fin et ont déjà été séparées lors de la cartographie (DPV 372). Dans cette bande de 25 km de large qui s'étend en arc de cercle sur 150 km de long, des basses-terres du Saint-Laurent jusqu'à 60 km au sud du lac Saint-Jean, les roches basiques, de composition tholéitique d'après les analyses de Pyke, sont plus abondantes à l'est et les roches acides et gneissiques à l'ouest.

Ce complexe comprend plusieurs types de roches que l'on observe habituellement dans les ensembles ophiolitiques. Le style tectonique et la répartition des faciès suggèrent l'existence de failles inverses ou de charriage à pendage vers l'est faisant se superposer des tranches distinctes des roches de ce complexe qui pourraient être des lambeaux de croûte océanique ou de bassin marginal. Les roches basiques étant plus denses que les autres on peut reconnaître leurs influences sur la carte gravimétrique (fig. 14).

CORRELATION ET METAMORPHISMES

A l'exception du groupe de Montauban et du complexe de La Bostonnais tous les autres groupes de la région possèdent des roches charnockitiques, c'est-à-dire, des roches qui se sont formées dans les conditions de stabilité de l'orthopyroxène. Sont-elles toutes de même âges? Pour le savoir il faut d'abord connaître l'âge relatif des roches affectées.

Corrélation des différents groupes métasédimentaires

Dans l'ordre chronologique le complexe de Chapeau de Paille semble bien être le groupe le plus ancien et avoir servi de socle aux dépôts de plate-forme plus jeunes (Wynne-Edwards, 1972).

Les groupes de Grenville et de Mékinac reposent tous deux sur le même socle. Leurs relations respectives sur le terrain ne sont pas connues et les quelques itinéraires effectuées dans la partie sud-ouest de la région ne sont pas suffisants pour en tracer les limites précises; notons cependant les grandes différences entre ces deux groupes; séquence volcanique (métagabbro) à la base de celui de Mékinac, sédiments argileux (gneiss à sillimanite) à la base de celui de Grenville, puis sédimentation mature détritique ou carbonatée de ce dernier que l'on ne rencontre qu'occasionnellement dans les niveaux B et C de celui de Mékinac etc. Deux solutions logiques sont possibles:

1. Le groupe de Mékinac est antérieur au groupe de Grenville qui lui est discordant.
2. Le groupe de Mékinac est au moins en partie l'équivalent latéral de celui de Grenville et représente une séquence de bord de marge continentale nord-sud avec volcanisme initial de séparation (M_1), dépôts continentaux (niveau A)? puis de plate-forme (niveau B) et enfin profonds (M_4). Une sédimentation d'aulacogène est encore possible pour le groupe de Grenville (Baer, 1976b).

Le groupe de Saint-Félicien est mal connu, et peut donc difficilement être corrélé à ceux de Grenville et de Mékinac. Il est possible aussi qu'il soit discordant sur ces deux derniers et que les roches basiques (gabbro, anorthosite) que l'on observe près du contact représentent une ancienne ligne de

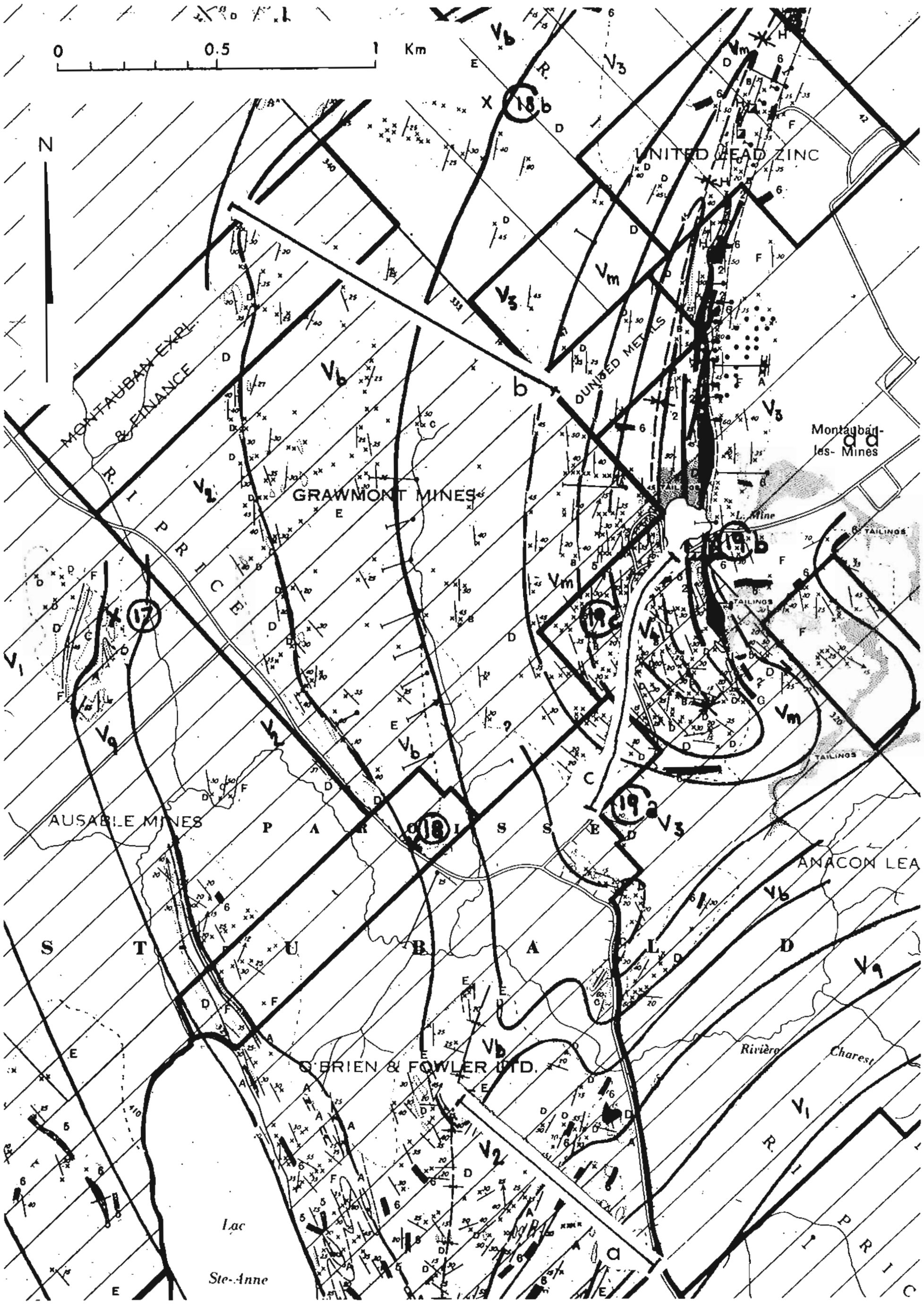

Figure 11. Fragment de la carte géologique de Smith (1956) avec interprétation photostéréoscopique et emplacement des coupes-types a, b, c.

suture, mais si tel est le cas elle serait très ancienne, hudsonienne? (tracés sinueux) et reprise par une autre orogénie majeure (foliation et métamorphisme discordants par rapport à cette ligne est-ouest).

Les roches du groupe de Montauban bien qu'entièrement marines ne semblent pas s'être formées loin du continent. Celles du groupe de La Bostonnais par leur composition sodique et basique en général et par le mélange intime roches massives-roches gneissiques semblent bien représenter l'association typique de la couche 3 d'une croûte océanique (cf. gabbro du complexe de Baie of Islands, Williams et Stevens, 1974) et de dépôts de profondeur et peut-être aussi de talus continental (flysch) lors d'une collision (Dewey, 1976; Al-Shanti et Mitchell, 1976).

S'il s'agit de lambeaux de croûte océanique, on devrait trouver à l'est du groupe de La Bostonnais des groupes métasédimentaires totalement différents. Cela correspond bien aux observations: les épais dépôts calcosilicatés, des groupes de Grenville et de Mékinac et aussi des Adirondacks, n'existent pas à l'est où on observe par contre plusieurs niveaux de conglomérat, et le groupe des Martres (Rondot et Marleau, 1977) se serait déposé sur un autre continent sous climat aride.

Charnockitisation

Le métamorphisme de cette partie du Grenville est donc complexe et en liaison, pour chaque endroit, avec son histoire tectonique. Sur la figure 4, on remarque que les seules isogrades qui on pu être isolés sont ceux de l'hypersthène et de la muscovite. Tous les deux semblent en discordance sur un équilibre antérieur comme ils le sont à travers les groupes métasédimentaires. Par ailleurs, la sillimanite semble stable dans toute la région.

Les roches métamorphiques à pyroxènes sont très particulières et ont très bien été décrites par Th. Holland (1908) aux Indes. Ces roches de teinte verte contiennent habituellement, en plus des deux pyroxènes de la hornblende et un peu de biotite. Les roches charnockitiques, de par leur composition minéralogique, se sont formées à haute température et donc bien à l'intérieur de la zone d'anatexie (Winkler, 1976). Elles sont, soit gneissiques (migmatite charnockitique), soit massives (charnockite, opdalite etc). Les roches massives peuvent s'éloigner sous forme de diapir (Schijver, 1975) et quitter ce que j'appellerai le front de charnockitisation.

Ce front est en fait une zone qui peut-être large (plusieurs kilomètres), surtout s'il est subhorizontal, où les transformations sont progressives. La première est associée à l'apparition de la teinte verte avant celle des pyroxènes. Le contact est brusque à l'échelle de l'échantillon (fig. 15, la foliation et la paragenèse à biotite grenat sont détruites), mais très irrégulier sous forme d'apophyse comme l'avait déjà décrit Holland (Rondot, 1964, DPV 372, p. 52). A l'intérieur de la zone charnockitique on note parfois autour des enclaves de roches basiques (pyribolite) une zone d'un centimètre d'épaisseur environ où la hornblende est presque entièrement

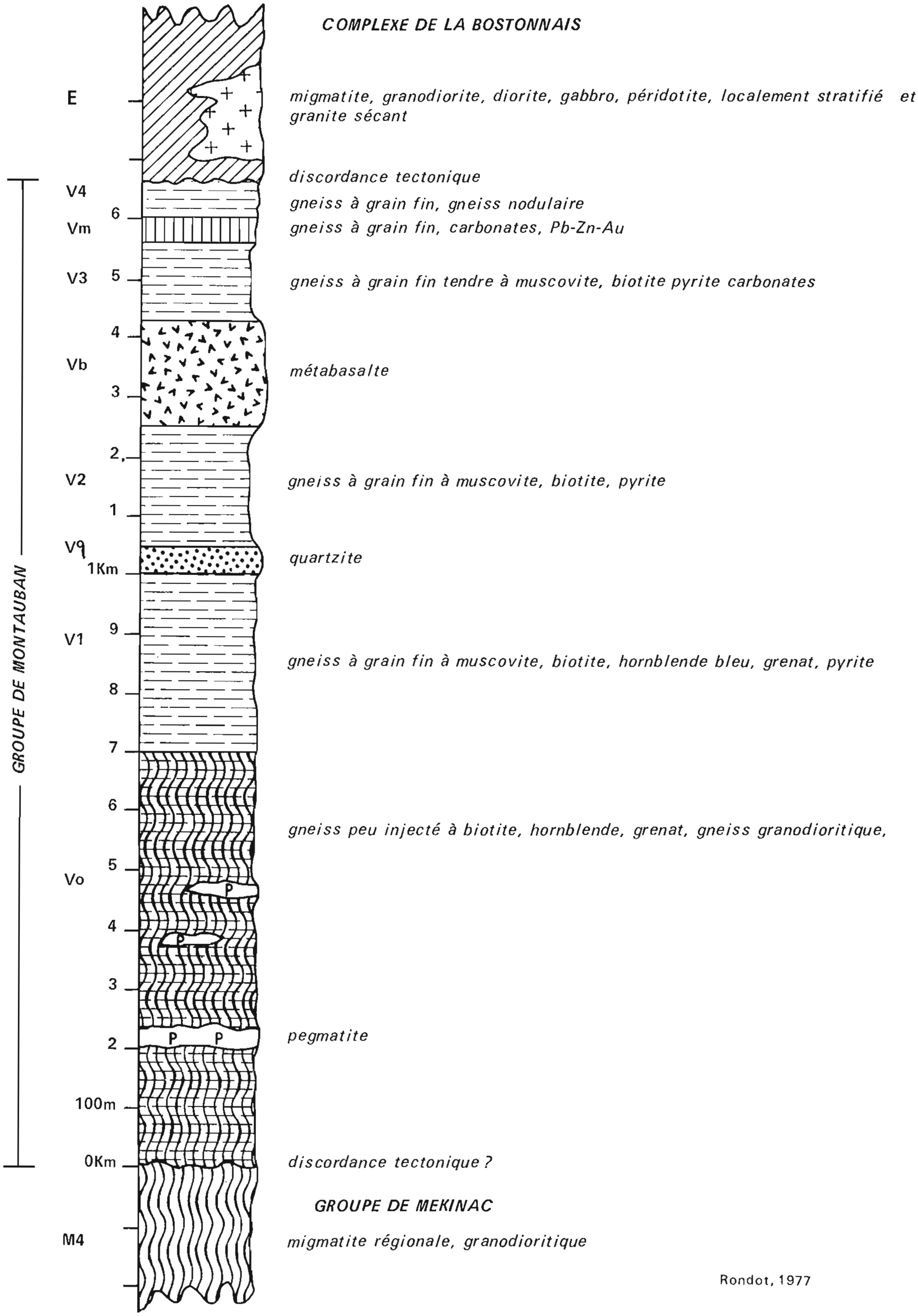

Figure 12. Echelle stratigraphique du groupe de Montauban.

Figure 13a.

Complexe de La Bostonnais. Affleurement à la sortie nord du village de La Croche. Diorite à grain moyen à grossier dans des gneiss à grain fin bien lités. (GSC 203364-D)

Figure 13b.

Détail de la figure précédente. (GSC 203364-E)

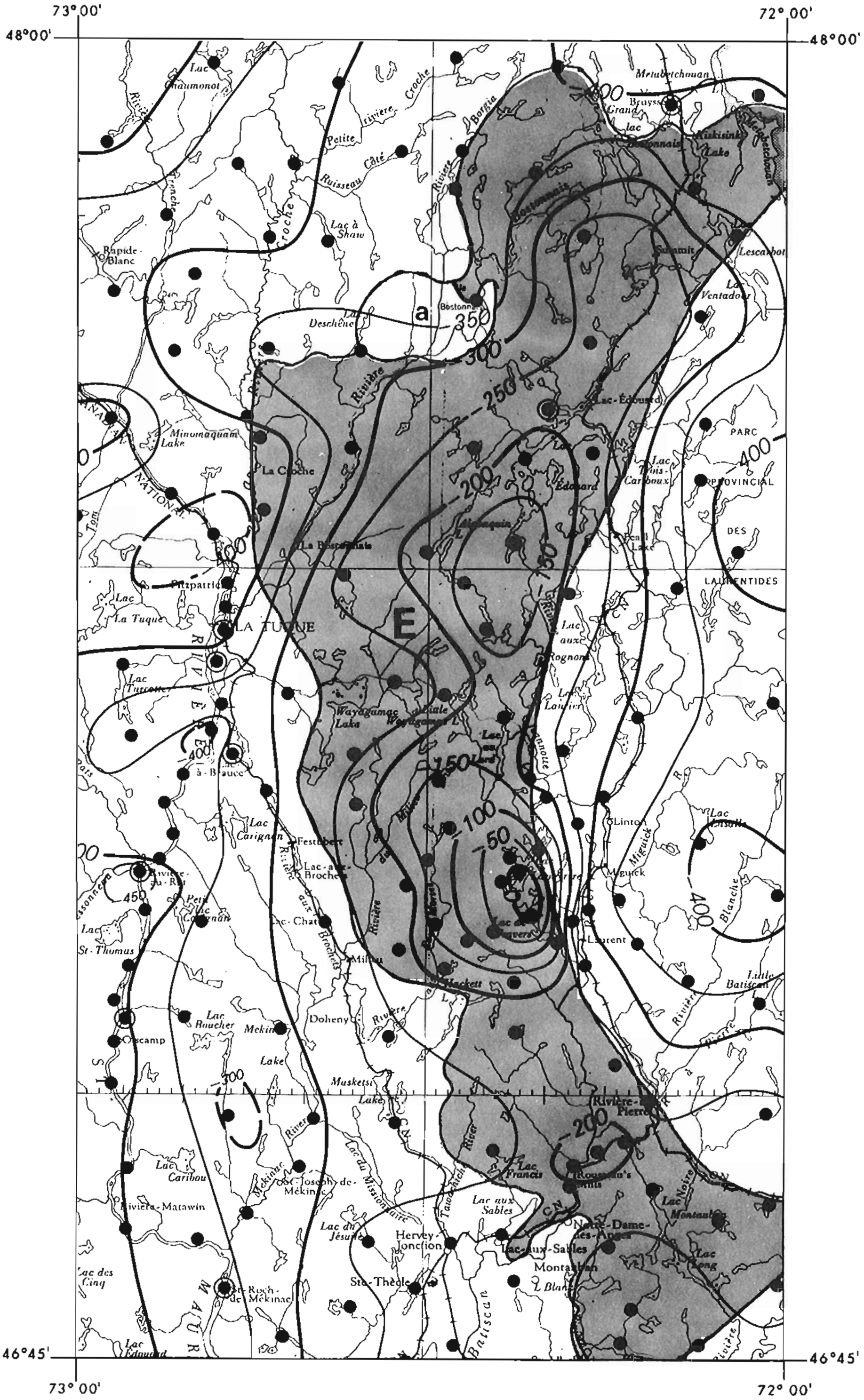

Figure 14. Fragment de la carte gravimétrique de la région de Trois-Rivières n° 135. En gris, roches du complexe de La Bostonnais, a – anorthosite de Langelier.

transformée en pyroxène au contact de la roche charnockitique felsique. Schrijver (1973, 1975) a observé des phénomènes semblables au sud-ouest de notre région.

La roche à saphirine, presque au coeur de la plus importante zone de migmatite charnockitique, au milieu de la coupe-type du niveau b, a été étudiée par Fergusson (1974) sous la directive de R.K. Herd. D'après Ferguson les conditions locales de métamorphismes était: pression totale entre 5 et 9 kb*, température environ 7 à 800°C et pression du fluide faible mais variable, 0 à 2,5 kb. La limite de 9 kb est due à l'absence de l'association: hyperstène, grenat, augite, plagioclase, quartz. Mais on peut dire aussi qu'elle est plus faible que 8 kb du fait de l'absence de l'association saphirine – quartz (Winkler, 1976, p. 267). D'autre part, la composition de la saphirine donnée par Fergusson (1974, p. 49), reportée sur le diagramme de Bishop et Newton (1974, p. 515) indiquerait des conditions de plus basse pression et plus haute température que celle de la saphirine du Fiskenaeset (Herd, 1973). La saphirine du Saint-Maurice est plus pauvre en SiO_2 et MgO et plus riche en oxide de fer. On peut en estimer une température de formation de 900°C pour une pression de 5 kb, assez semblable aux estimations de Hermans et al. (1976) pour un gisement de saphirine du sud-ouest de la Norvège.

Il est à noter qu'en dehors du front de charnockitisation le faciès amphibolite n'est pas uniforme. En effet, on observe dans la partie sud-ouest de la région, et non loin des roches charnockitiques, des migmatites où la muscovite est stable en présence de quartz comme cela avait déjà été observé plus à l'est (Wynne-Edwards et al., 1966).

En conclusion, on peut dire que le phénomène de charnockitisation suppose un métamorphisme préalable qui a chassé l'eau et a donné aux roches une certaine rigidité permettant la conservation de certaines structures primaires (métagabbro) et un faible déplacement des masses rocheuses (roche charnockitique massive peu éloignée du front de charnockitisation). Un premier métamorphisme n'aurait donné à l'extrême que des roches intrusives (granite d'anatexie) mais non dépourvues d'eau.

Le front de charnockitisation est ici indépendant de l'ordre stratigraphique et donc postérieur aux différentes unités sauf au groupe de Montauban et au complexe de La Bostonnais avec lesquels il pourrait être relié d'une certaine manière.

Gneiss à muscovite *(groupe de Montauban)*

Les roches du groupe de Montauban offrent de très belles textures primaires elles ont pourtant subi plusieurs périodes de déformation et c'est à une des dernières (plis déversé vers l'ouest si l'on se fie à l'alongement nord-nord-ouest de cette bande) que semble due l'apparition de la muscovite et aussi celle des gneiss nodulaires.

D'ouest en est, l'apparition de la muscovite dans les roches du groupe de Montauban se fait juste avant la disparition progressive des injections et l'intercalation d'une zone de pegmatites blanches (DPV 373). A Montauban, il n'y a pas d'injection sauf celles reliées aux intrusions proches. Dans les gneiss à biotite à grain fin, la muscovite et le quartz sont en équilibre avec le microcline, mais la sillimanite n'apparaît que dans les gneiss nodulaires sous forme de fribolite. Les nodules, ovoïdes, blanc laiteux ne contiennent

* 1kb = 0.1 GPa

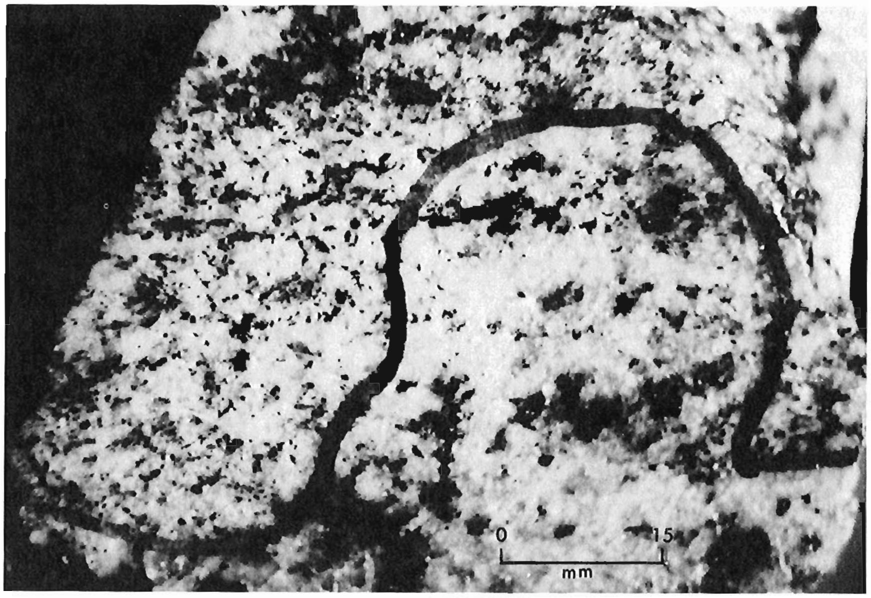

Figure 15.

Apophyse de roche charnockitique dans le gneiss à biotite et grenat, échantillon pris à 7 km du barrage de Beaumont sur le Saint-Maurice. (GSC 203364-F)

qu'une mosaïque de quartz avec des paquets enchevêtrés de fibrolite. La matrice est à quartz, microcline, biotite, muscovite avec ou sans plagioclase. L'origine de ces gneiss nodulaires observés depuis près d'un siècle ne semble pas être connue de façon définitive.

Le minerai de Montauban offre une grande variété de minéraux, aussi, K. Stamatelopoulou-Seymour considérant que la recristallisation métamorphique a été complète, utilise plusieurs équilibres minéralogiques dont celui de la blende comme géobaromètre et obtient des limites de 630 à 600°C de température et 3,5 à 4,5 kb de pression. Si nous reportons ces résultats et ceux de Ferguson sur le diagramme pression-température de Winkler (1976) nous remarquons, en utilisant la partie haute température basse pression de la marge de possibilité des conditions de P et T obtenues par Ferguson, que les conditions de métamorphisme corresponderaient au type C de métamorphisme régional de Winkler. Il faut noter que l'estimation de Ferguson est la zone d'anatexie de Winkler.

Métamorphismes et tectoniques

Les observations effectuées dans la région du Saint-Maurice nous indiquent qu'il y a eu au moins trois métamorphismes. Le dernier probablement dû à une déformation de type collision continentale n'a en général que peu affecté les roches si ce n'est par une cataclase et un rétromorphisme assez général et des injections alaskitiques en particulier le long des failles majeures. La datation au potassium-argon donnerait autour de 950 Ma (900 pour la pseudotachylite de la faille du Saint-Maurice, (Philpotts et Miller, 1963)).

L'avant dernier métamorphisme est celui qui a conduit à la formation des charnockites et peut-être aussi de la zone à muscovite de Montauban. S'il n'y a qu'une seule période de formation de ces charnockites, leur dispersion dans une zone assez large, plusieurs centaines de kilomètres, et de direction nord-est semble erratique. Ces zones de roches charnockitiques correspondraient à des dômes de chaleur dont les résultats seraient différents d'un endroit à l'autre. Au centre de la région nous avons vu qu'il n'y avait pratiquement que des migmatites charnockitiques. Dans le nord-est de la région par contre (groupe des Weymontachingues sur la figure 4) ces dernières sont accompagnées de nombreuses intrusions comme on l'a aussi observé dans la région de Charlevoix (Rondot, 1972b, p. 143). Les intrusions à grain grossier avec phénocristaux de 1 à 2 cm sont d'abord acides (charnockites, monzonite quartzique à pyroxènes) et syncinématiques, puis en massifs de composition intermédiaire (mangérite, jotunite) et discordants; ces derniers ayant migrés en dehors de la zone à hypersthènes dans les migmatites à biotite hornblende. Cela correspond bien, comme dans Charlevoix à une suite palingénétique où, à cause du manque d'eau les pegmatites sont très rares, mais la température étant d'autant plus élevée a provoqué la fusion de roche acide à intermédiaire.

Si l'on remonte encore dans le temps, on s'aperçoit que la région du Saint-Maurice a un comportement différent de celle de Charlevoix par exemple. Chacune des régions possède une petite masse d'anorthosite (fig. 16). L'une est très déformée, l'autre très fraîche. Celle de Langelier (Ryder, 1974; Baer, 1976a) dans la région du Saint-Maurice a subi une importante déformation post-magmatique comme d'ailleurs aussi l'anorthosite de Morin (Martignole et Schrijver, 1970 et Martignole, 1975) et celle du lac Saint-Jean (Berrangé, 1962; Rondot, 1963; Kehlenbleck, 1972). L'anorthosite de Langelier est entièrement granulée et certains affleurements, très foliés, ont l'apparence de roches sédimentaires. Cette foliation, uniquement mécanique, la roche contient 99% de plagioclase An 50-55, est nette sur les photographies aériennes et discordante par rapport au contact avec les roches encaissantes. Vers le centre de la masse la proportion de minéraux ferromagnésiens est plus importante. Il s'agit d'aglomérats allongés d'olivine, atteignant une dizaine de centimètre de diamètre sur un mètre de long, entouré d'hypersthène, hornblende, spinel, opaques etc.

Au contact de ces amas de minéraux ferromagnésiens le plagioclase est appauvri en sodium. Les roches charnockitiques qui entourent ce massif sont à grain fin et très fraîches et contiennent des enclaves d'anorthosite. Elles ont cependant été en partie affectées par les dernières orogénies, matérialisées surtout par des injections et pegmatites alaskitiques et des cataclases.

L'anorthosite de Saint-Urbain, possède par contraste des caractères primaires bien conservés. Cette anorthosite à hypersthène se présente sous plusieurs faciès de composition légèrement différente (Mawdsley, 1927; Rondot, 1972a); anorthosite rubannée en enclave dans l'anorthosite à labrador à grain moyen, dykes d'anorthosites pegmatitiques à andésine etc. Comme l'anorthosite de Langelier, celle de Saint-Urbain est antérieure aux roches charnockitiques qui en contiennent des enclaves et la recoupe sous forme de dykes. Les déformations tardives sont peu importantes mais présentes.

Ces deux masses anorthositiques ont donc subi également une période de déformation (950 Ma) et une période de charnockitisation (1 250 Ma) mais présentent une histoire antérieure différente: une phase de déformation supplémentaire pour l'anorthosite de Langelier. Ces deux massifs sont, ou d'âge très différent, ou mis en place dans des continents différents reliés maintenant par la suture du complexe de La Bostonnais.

CONCLUSION

1. La diversité des groupes et complexes de la région du Saint-Maurice suppose une histoire tectonique et métamorphique en plusieurs étapes, la première est la formation d'un socle de roches granodioritiques peu diversifiées; complexe de Chapeau de Paille, sur lequel repose des groupes métasédimentaires de plate-forme et de marge continentale; groupes de Grenville et de Mékinac auxquels on pourrait peut-être ajouter le groupe de Saint-Félicien si ce dernier n'est pas séparé des autres par une suture tectonique est-ouest. Le groupe métavolcanique de Montauban et le complexe de La Bostonnais apportent un élément nouveau et séparent des domaines différents de l'écorce terrestre.

2. Les isogrades de métamorphisme les plus évidents, hypersthène et muscovite, sont discordants sur le patron stratigraphique et semblent aussi l'être sur un héritage métamorphique antérieur.

3. La tectonique est complexe et différente d'un endroit à un autre. Les zones les moins déformées, dans le groupe de Montauban par exemple, indiquent au moins trois phases de déformation. Il semble cependant qu'une seule phase soit à l'origine de plis couchés alors que les roches n'étaient pas encore rigides. Pour essayer de retrouver les différentes phases, on peut procéder par ordre chronologique inverse.

4. Les failles, les injections alaskitiques, le rétromorphisme et en partie la granulation semblent appartenir à un même cycle qui donnerait les dates au potassium-argon de 1 100 Ma, particulière à la province de Grenville.

5. La charnockitisation est un phénomène plus ancien auquel on doit les dates au potassium-argon de 1 205 Ma sur une monzonite à pyroxènes du mont Tremblant à quelques km au sud-ouest de notre région (Wynne-Edwards et al., 1966), 1 280 Ma au Rb-Sr sur du matériel provenant de roches charnockitisées dans Charlevoix (Rondot, 1971) etc. La charnockitisation est le métamorphisme de roches déjà métamorphiques donc pauvres en eau. Elle correspond au métamorphisme ordinaire avec ses migmatites et ses granites d'anatexie, mais avec des conditions différentes: plus haute température de formation, presqu'absence de pegmatite et grain plus fin dû au manque d'eau, faible déplacement des masses palingénétiques formées, de sorte que l'on peut observer presque sur place; migmatites charnockitiques, intrusions acides syncinématiques et masses discordantes de composition intermédiaire dont la minéralogie rappelle celle de roches du socle fondues à haute température (Floran et al., 1976). La limite de ce métamorphisme que l'on peut appeler front de charnockitisation est large et laisse les roches encaissantes peu déformées.

6. Une zone de suture, de direction nord-sud dans notre région, matérialisée par le groupe de Montauban et le complexe de La Bostonnais, pourrait être reliée à la charnockitisation par les zones de subduction, sources de chaleur, qu'elle suppose. Les roches charnockitiques, en partie syncinématiques, correspondent à une période de réactivation lors de compression tandis que la mise en place des anorthosites plus anciennes, en partie de type diapirique et souvent non déformées après leur cristallisation primaire, correspondraient plutôt à une période de tension ou du moins de calme tectonique.

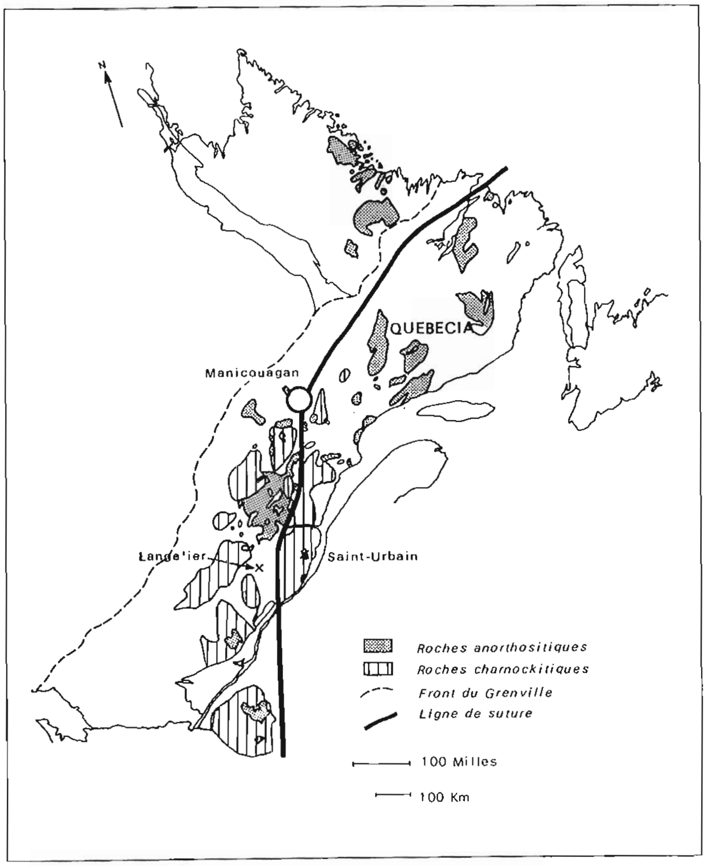

Figure 16. Ligne de suture séparant les continents Laurentia et Quebecia. Disposition des roches anorthositiques et charnockitiques dans la province de Grenville modifiée de Emslie (1975).

Cette suture se poursuivrait jusqu'à l'astroblème de Manicouagan où Kish (1968) a observé d'épaisses formations basiques à ultrabasiques stratifiées (complexe du lac Raudot) et où une anomalie gravimétrique prononcée (Thomas et Tanner, 1975; Gibb et Thomas, 1976) se prolonge jusqu'au-delà du lac Ashuanipi et au sud de la zone de rift de Seal Lake (Burke et al., 1977). Au sud du Québec cette suture longerait la limite est des Adirondacks qui présentent les mêmes caractères stratigraphiques que notre région (fig. 16). Ainsi aux noms de Laurentia et Grenvillia (Irving et al., 1974), on pourrait ajouter Quebecia pour désigner cette partie de continent à l'est de cette ligne de suture et sur laquelle se trouve Québec.

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ANNEXE I

Coupe type du "niveau B" du groupe de Mékinac

Coupe type de la partie inférieure de l'unité M_3, "niveau B" du groupe de Mékinac (72°45,3'W, 46°51,4'N, fig. 6, site 4 et fig. 7 et 8) à 6,5 km au nord-ouest de Saint-Roch de Mékinac le long de la route de La Tuque. Série renversée, pendage vers le nord-est (description d'après la coupe faite par Joël Brun, Rondot 1976b).

Coupe de la falaise à saphirine du Saint-Maurice

UNITES (du sommet à la base, NW au SE)		Epaisseur (m)
M_3 supérieure		
1	- migmatite charnockitique avec lentilles de roches calcosilicaté (zone plissée)	>10
M_3 inférieure, "niveau B" (total 196 m)		
19	- roche à calcite rose à gris pâle avec enclaves de gneiss gris noirâtre à grenat	12
18	- roche calcosilicatée grise à quartz	3 à 4
17	- roche à calcite à enclave (cf. 19)	2 à 3
16	- migmatite charnockitique verte avec, vers le milieu, lentilles de roche calcosilicatée	20
15	- quartzite gris à grain grossier vers la base	0 à 5
14	- migmatite charnockitique verte, bien litée vers le milieu	36
13	- roche calcosilicatée grise et quartzite	12
12	- migmatite charnockitique verte	22
11	- gneiss à saphirine	0 à 1
10	- migmatite charnockitique verte	4
9	- roche grise à carbonate et calcosilicatée	6
8	- migmatite charnockitique, bien litée au sommet, amas irrégulier de calcitite le long d'une discordance à la base (fig. 7)	25
7	- migmatite charnockitique verte lentille de roche calcosilicatée vers le milieu	8
6	- roche grise à carbonate et calcosilicatée	2
5	- migmatite charnockitique verte	8
4	- quartzite, roche calcosilicatée grise et schiste à biotite avec nodules de quartz discordant à la base	3
3	- migmatite charnockitique avec passées de quartzite vers le sommet	10
2	- roche calcosilicatée, schiste à biotite, quartzite gris et nodules de diopside à phlogopite (fig. 8)	2
1	- quartzite et gneiss quartzeux à grenat, bien stratifiés (couches de 30 cm), passées de roche calcosilicatée	14
M_2 supérieure, migmatite charnockitique verte		>3

ANNEXE II

Coupe type du "niveau C" du groupe de Mékinac

Coupe type de la partie inférieure de l'unité M_4, "niveau C" du groupe de Mékinac (72°37,7'W, 46°45,5'N, fig. 9, site 12) à 6 km à l'ouest-nord-ouest de Saint-Tite. Série normale à pendage de 40° vers le nord-est. Un chemin non carrossable se rendant au lac Silhouette, donne accès à cette coupe. Description sommaire d'après deux itinéraires.

Coupe du lac Silhouette

UNITES (du sommet à la base, NE au SW)	Epaisseur (m)
M_4 supérieure	
- migmatite granodioritique	>10
M_4 inférieure, "niveau C" (total environ 450 m)	
- migmatite quartzeuse à biotite, grenat, graphite; gneiss rouillé	50
- migmatite plagioclasique vert cassonade à grain fin à moyen, passage riche en pyroxène, hornblende, biotite ou en diopside, microcline	200
- gneiss quartzeux lité (biotite)	20
- migmatite plagioclasique à hornblende, pyroxène	30
- gneiss à silimanite	30
- gneiss à grain fin à biotite pyroxène, gneiss graphiteux	20
- migmatite (hétérogène) et gneiss (régulièrement lité) très variés: quartzeux à feldspath vert et grenat mauve (sillimanite), à minéraux calcosilicatés et carbonatés, à feldspath vert, rouillés etc.	100
M_3 supérieure	
- migmatite charnockitique verte, un niveau de gneiss à scapolite, diopside, sphène	>100

ANNEXE III

Coupe type du groupe de Montauban, moitié supérieure

Partie C: coupe type des unités V_m et V_4 (72°21,1'W, 46°49,3'N, fig. 12 coupe C), route menant au village de Montauban-les-Mines. 1 km au sud-ouest du village, flanc ouest du synclinal. Pendage de 20° vers le nord-est.

Coupe de la route

UNITES (du sommet à la base, NE au SW)			Epaisseur (m)
		Lacune stratigraphique	
V_4	3	- couches et lentilles dans 2, décimétriques à métriques de gneiss blanc à grain moyen, à quartz, plagioclase, microcline, muscovite et nodules de quartz et fibrolite avec ou sans muscovite.	
	2	- gneiss gris pâle à grain fin à moyen, à quartz, feldspath, muscovite, biotite, avec lits millimétriques, à muscovite et phlogopite bien cristallisées.	30
	1	- gneiss gris à grain très fin à quartz, feldspath, biotite, opaques finement stratifiés; les lits, millimétriques à centimétriques, contenant des proportions relatives variées des minéraux ci-dessus.	23,5
V_m	3	- gneiss rubannés gris, mauve et pâle, à grain fin à moyen à quartz, plagioclase, biotite avec lentilles de carbonate contenant quelques sulfures et une couche de 20 cm d'amphibolite à hornblende, biotite, épidote, chlorite presque dépourvues de plagioclase.	4,5
	2	- gneiss gris pâle massif à bien stratifié et granoclassé avec minces lentilles de carbonate.	30
	1	- gneiss à grain fin rouillé, tendre; couche de gneiss à hornblende, grenat au sommet; 10 m sans affleurement à la base.	15
V_3		- gneiss à grain fin rubanné, à biotite, opaques, (hornblende), muscovite discordante; débit en plaquettes; niveaux à grenat; quelques lentilles à carbonate.	20
		- pas d'affleurement.	6
		- gneiss à grain fin à biotite et opaques.	1
		- pas d'affleurement.	4
		- gneiss à grain fin, petits lits d'amphibolite et de quartz	1
		- pas d'affleurement.	2
		- gneiss quartzofeldspathique rubanné, mauve à vert à grain fin; lits de 2 à 5 mm de composition variées (carbonate, opaques, graphite, biotite; carbonate, hornblende, biotite, sulfure; hornblende, apatite, biotite), lentille de quartz.	6

ANNEXE III (suite)

Partie B: coupe type des unités V_b et V_3 (72°21,8'W, 46°50'N, fig. 12, coupe b) de direction W 30°N et passant par le sommet de la colline de métabasalte, contact V_3, V_b à 1,5 km à l'ouest-nord-ouest de Montauban. Pendage de 25° vers l'est.

Coupe de la colline de métabasalte

UNITES (du sommet à la base, SE au NW)		Epaisseur (m)
V_3	- gneiss variés, lités, à grain fin; lits pâles à biotite-muscovite, gris à biotite-hornblende, sombre à hornblende; affleurements très isolés.	130
V_b	- amphibolite à grain fin, litée au sommet (coussinets de la base au sommet) (130 à 180 m).	150
V_2	- gneiss rouillé à grain fin à quartz, plagioclase, biotite, grenat.	60

Partie A: coupe type des unités V_q et V_2 (72°21,3'W, 46°48,4'N, fig. 12, coupe A) de direction W 40 N à partir d'un coude de la route Saint-Ubald – lac au Sable, en direction du sommet d'une petite colline. Pendages variables vers le nord-ouest.

Coupe du lac Sainte-Anne

UNITES (du sommet à la base, NW au SE)		Epaisseur (m)
V_b	- amphibolite à coussinets écrasés.	25
V_2	- gneiss à grain fin, lité, à biotite (grenat, hornblende); passages à muscovite; (affleurements dispersés); quelques pegmatites.	200
V_q	- quartzite, quartzite feldspathique, gneiss quartzeux, pegmatite vers le sommet (affleurements discontinus).	50
V_1	- gneiss rubanné, amphibolite litée.	10

METAMORPHISM IN CENTRAL GRENVILLE PROVINCE

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Sharma, K.N.M., Clark, T., Franconi, A., and Laurin, A.F., Metamorphism in central Grenville Province; in Metamorphism in the Canadian Shield, Geol. Surv. Can., Paper 78-10, p. 353-356, 1978.

Abstract

The region north of the St. Lawrence River between 63°00'W and 72°30'W is underlain principally by pre-Elsonian gneiss and by intrusive Elsonian (1400 Ma) anorthositic, charnockitic, and granitic plutons. The gneiss, which has been subjected to several episodes of deformation and metamorphism, has been grouped into three major rock units: (1) grey gneiss, consisting of mineral assemblages of amphibolite facies grade; (2) charnockitic gneiss, containing orthopyroxene and clinopyroxene ± hornblende ± biotite, some of which are granulite facies equivalents of (1); (3) paragneiss, characterized by the assemblage plagioclase-potash feldspar-quartz-biotite ± garnet ± sillimanite ± muscovite ± hornblende ± graphite ± cordierite (rare). Granulite facies gneiss is abundant around the anorthosites in the western part of the area, but is rare around the eastern anorthosites. This distribution is probably a result of retrograde metamorphism in the east. Cataclastic deformation and recrystallization affected almost all plutonic rocks to varying degrees during the Grenville Orogeny.

Résumé

Le sous-sol de la région située au nord du fleuve Saint-Laurent, entre 63° 00'W et 72° 30'W comprend principalement des gneiss pré-elsoniens, et des plutons intrusifs de caractère anorthositique, charnockitique et granitique, datant de l'Elsonien (1 400 Ma). Le gneiss, qui a subi plusieurs épisodes de déformation et de métamorphisme, regroupe trois principales unités lithostratigraphiques: (1) un gneiss gris, constitué d'assemblages minéraux du faciès des amphibolites; (2) un gneiss charnockitique, constitué des minéraux suivants: orthopyroxène et clinopyroxène ± hornblende ± biotite, dont certains sont les équivalents de (1) dans le faciès des granulites; (3) un paragneiss caractérisé par l'assemblage plagioclase ± feldspath potassique ± quartz ± biotite ± grenat ± sillimanite ± muscovite ± hornblende ± graphite ± cordiérite (rare). Le gneiss du faciès des granulites abonde aux alentours des anorthosites du secteur ouest, mais est rare à proximité des anorthosites du secteur est. Cette distribution résulte probablement d'un métamorphisme régressif à l'est. Des processus de déformation cataclastique et de recristallisation ont modifié toutes les roches plutoniques pendant l'orogenèse du Grenvillien.

INTRODUCTION

The crystalline rocks that underlie the area shown in Figure 1 comprise part of the metamorphic terrane of the Grenville Structural Province of the Canadian Shield. The Grenville Orogeny (950 ± 150 Ma in age) represents the last metamorphic episode affecting the rocks of the region. The rocks of the province are in general characterized by a higher grade of metamorphism compared with the rocks of the adjacent Superior Province, and by the presence of several large anorthosite massifs and related rocks of the charnockitic suite. The emplacement of the anorthosites, plutonic charnockitic rocks, and some granitoid rocks represents the major intrusive activity of Elsonian time (1400 Ma ago). Postorogenic rocks comprise mainly unmetamorphosed granite-pegmatite, diabase, and carbonatite dykes. In general, the grade of metamorphism ranges from amphibolite to granulite facies; greenschist facies occurs in some areas underlain by paragneiss of the Wakeham Bay Group and in areas where retrograde metamorphism has taken place.

To facilitate the compilation of the large volume of data collected in the reconnaissance mapping program of the Grenville Project by the Quebec Department of Natural Resources, it was necessary to group the various types of gneiss into three major categories based mainly on their textural, structural, and mineralogical characteristics:

1) Grey gneiss, banded grey gneiss, and their migmatitic equivalents;
2) Green charnockitic gneiss, mostly concentrated in the vicinity of anorthosite massifs, and plutonic charnockitic rocks; and
3) Paragneiss, including quartzite, crystalline limestone, and calc-silicate rock.

The gneiss has been intruded by various plutonic rocks consisting mostly of anorthositic and associated charnockitic rocks, and by granitic, monzonitic and syenitic rocks.

Structural analyses indicate that the gneissic rocks have undergone several phases of deformation and metamorphism, whereas the intrusive rocks have undergone deformation and metamorphism only during the Grenville Orogeny. This orogeny produced a well defined retrograde metamorphism and numerous cataclastic features which are preserved in the anorthosites and associated charnockitic rocks.

METAMORPHISM

The mineralogy of the gneissic and intrusive rocks indicates that the whole area was affected by regional metamorphism. Most gneiss contains minerals diagnostic of the amphibolite facies, but in the vicinity of the anorthositic intrusions, granulite facies assemblages are more common. Indications of retrograde metamorphism have been observed in many places. A description of the mineral assemblages characteristic of each group of rocks follows.

Grey gneiss, banded grey gneiss, and migmatized equivalents

The grey gneiss has a uniform mineralogical composition, and is homogeneous, medium to coarse grained, and generally leucocratic with a granoblastic texture. Ferromagnesian minerals constitute less than 20 per cent of these rocks. A well developed foliation is defined by lenticular streaks of biotite and/or hornblende.

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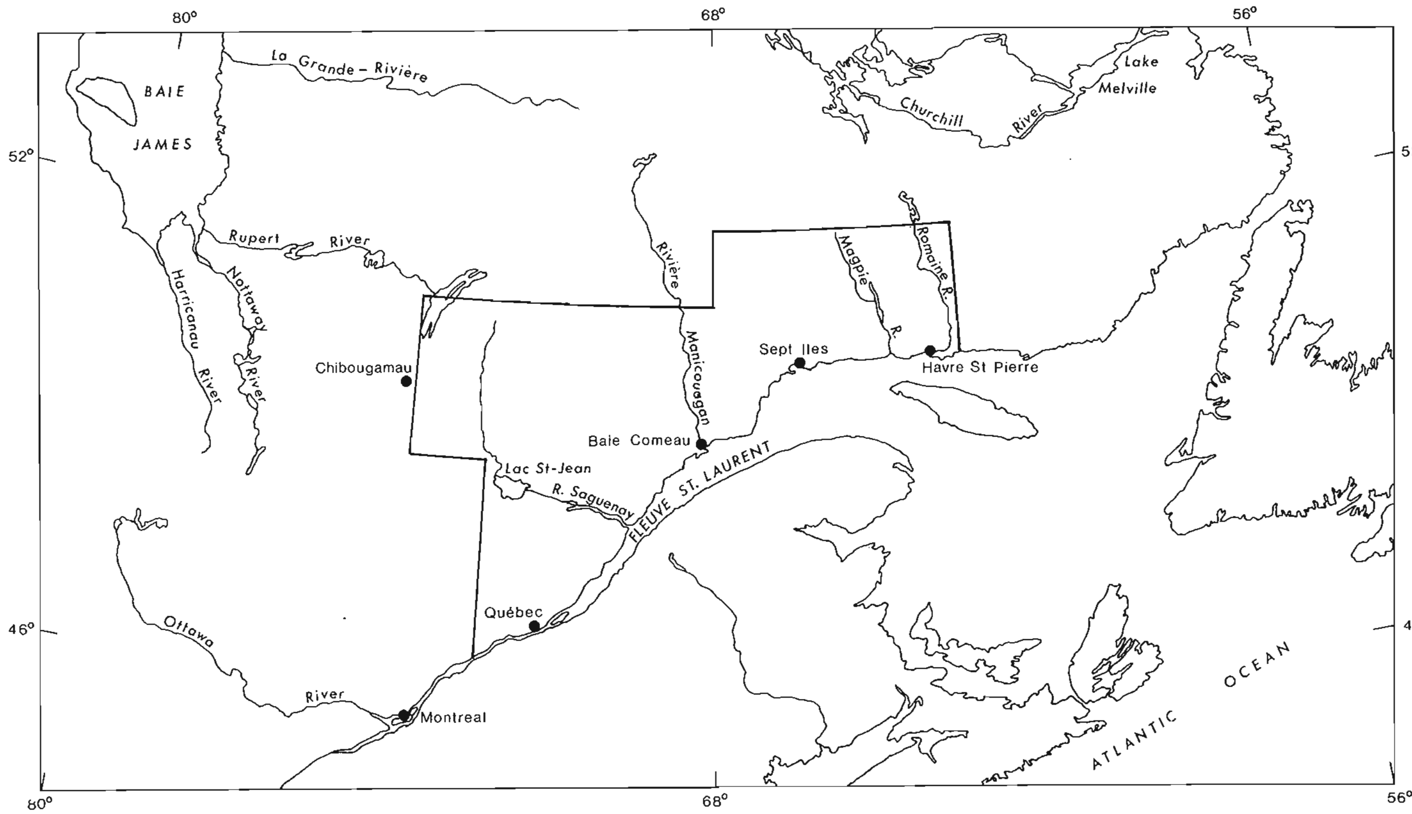

Figure 1. Map shows area of study in central Grenville Province.

The banded grey gneiss, sporadically developed and associated with the grey gneiss, possesses a laterally continuous compositional layering of quartzofeldspathic and mafic minerals. The bands vary in thickness from one centimetre to several metres and the grain size is generally finer than that of the typical grey gneiss. The grey and banded gneiss have been migmatized to varying degrees.

The grey gneiss, banded gneiss, and their migmatized equivalents, have a uniform mineralogical composition. The leucocratic minerals are plagioclase, quartz, and potash feldspar; the potash feldspar, which is predominantly microcline, may be absent. The ferromagnesian minerals are principally biotite and hornblende; garnet and/or clinopyroxene occur only sporadically. These minerals indicate the gneiss has experienced at least amphibolite facies metamorphism. In places, biotite and hornblende have been retrograded to chlorite and epidote. Where this effect is less marked, ovoid epidote grains have developed along the cleavage of biotite and hornblende; elsewhere, epidote occurs in well-defined crystals. Plagioclase is calcic oligoclase, or rarely, sodic andesine.

Charnockitic gneiss

This group of rocks includes gneiss that is characterized by light yellowish green to dark green fresh surfaces and distinctive white weathered surfaces. Weathered zones are yellowish brown and range from a few centimetres to several metres thick. Structurally and texturally, two principal types of charnockitic gneiss can be distinguished and mapped:

1) Fine to medium grained, homogeneous charnockitic gneiss, with a poor to well developed foliation similar to that observed in the grey gneiss, and

2) Heterogeneous, banded charnockitic gneiss, characterized by the alternation of quartzofeldspathic and mafic bands.

Feldspar, the principal constituent of the gneiss, is responsible for the green colour of the fresh rock. Quartz is present as yellowish or bluish grains, commonly elongated. The essential difference between the charnockitic gneiss and the grey gneiss is the presence of orthopyroxene and clinopyroxene which are associated with hornblende and biotite, and form small rusty spots on weathered surfaces. Thus, homogeneous charnockitic gneiss represents the granulite facies equivalents of the grey gneiss. Similarly, the banded charnockitic gneiss may represent either paragneiss or banded grey gneiss of the granulite facies.

The charnockitic gneiss, which mainly occurs near anorthosite massifs and associated rocks of the charnockitic suite, has been metamorphosed to granulite facies rank in one or both of the following ways:

1) At the depths corresponding to the emplacement of the anorthosites, which are deep-seated intrusions, the country rocks were already under the influence of metamorphic conditions at least approaching those of the amphibolite facies. The introduction of the anorthosites locally contributed in elevating the conditions of metamorphism to those of the granulite facies. Rocks beyond the influence of the anorthosites developed only mineral assemblages typical of the amphibolite facies.

2) The emplacement of the anorthosites and intrusive charnockitic rocks heated and dehydrated the surrounding gneiss. During the subsequent Grenville Orogeny the scarcity of water ($P_{H_2O} << P_s$) in the gneiss surrounding these intrusions favoured the development of orthopyroxene, whereas, in the group of rocks which did not undergo such heating, amphibole and biotite formed rather than orthopyroxene.

In the western part of the area high grade rocks are abundant around the anorthosites (e.g. the lac St. Jean anorthosite massif). In the eastern part of the area retrograde metamorphism has affected the anorthosite (e.g. the Havre St. Pierre, Romaine River and Magpie River massifs) and the surrounding gneiss. These anorthositic massifs were crushed, deformed, and extensively recrystallized. The gneiss has been retrograded to amphibolite facies, the earlier granulite facies assemblages are preserved in only a few places.

Charnockitic gneiss has a uniform mineralogical composition and contains plagioclase, potash feldspar and quartz. Ferromagnesian minerals do not exceed 25 per cent; orthopyroxene, an essential constituent, may be associated with one or more of the following: clinopyroxene, hornblende, and biotite. Accessory minerals include zircon, apatite, opaques, and allanite. However, the plagioclase and potash feldspar contents vary widely. The potash feldspar is generally microperthitic to mesoperthitic orthoclase, and the plagioclase in places is antiperthitic.

In conclusion, the charnockitic gneiss does not comprise a separate and distinct stratigraphic unit nor does it possess any unique textural and structural characters that would distinguish it from the other types of gneiss. Although it contains minerals typical of the granulite facies, the various textures and structures are identical to those of the amphibolite facies grey gneiss, banded grey gneiss, granitic gneiss, and paragneiss. Hence the charnockitic gneiss may be considered to represent the granulite facies equivalents of these rocks.

Paragneiss

These rocks are differentiated from the gneiss described previously by a much greater heterogeneity at outcrop scale, a well-developed and ubiquitous compositional layering, the presence of rocks of undoubted sedimentary origin, such as quartzite and crystalline limestone, and by the presence of alumino-silicates and graphite.

The paragneiss is similar to other paragneiss of the Grenville Supergroup. However, crystalline limestone is quite rare in the central Grenville Province compared with the southwestern part of the province. Crystalline limestone, because of its limited thickness and extension along strike, is of little use in stratigraphic correlation. The most common variety of paragneiss consists of the assemblage quartz-plagioclase-K feldspar-biotite ± hornblende ± garnet ± sillimanite ± muscovite ± graphite. Cordierite is very rare. The typical paragneiss is fine to medium grained; fresh surfaces are light to dark grey and weathered surfaces are rusty. The style of migmatization in the paragneiss is similar to that observed in the grey gneiss.

Plutonic rocks

The plutonic rocks of the area may be grouped into an anorthositic suite, a charnockitic suite, and a granitic suite. The anorthositic suite includes anorthosite, gabbroic anorthosite, anorthositic gabbro, gabbro, and mafic gabbro, as classified on the basis of ferromagnesian mineral content. These rocks form several large deep-seated massifs and numerous smaller intrusions. Primary minerals in these rocks include plagioclase, orthopyroxene, clinopyroxene, magnetite, ilmenite, olivine, zircon, and apatite. In places, these minerals show protoclasis; plagioclase and pyroxene crystals are bent and broken and are not recrystallized. The anorthosites also exhibit numerous features of cataclastic deformation acquired during the Grenville Orogeny. The fabric of these rocks ranges from slightly crushed and very coarse grained to well crushed and gneissic. The least deformed anorthosites are characterized by the presence of mortar texture around big plagioclase crystals. As the cataclasis becomes more pronounced, fewer of the original plagioclase phenocrysts can be identified. The crushed material shows recrystallization. The end product of cataclastic deformation is a white, granoblastic, equigranular rock with a typical sugary texture.

The cataclastic deformation and the accompanying recrystallization also affected the ferromagnesian minerals present in the anorthosite in a variety of ways. In the earliest stages of deformation development of hornblende, coronas formed around the pyroxenes; in the advanced stages of crushing and recrystallization symplectic textures developed, and the pyroxenes and plagioclase were completely transformed to hornblende, biotite, and garnet. The development of hornblende from orthopyroxene and plagioclase may be accompanied by very small grains of quartz in the symplectite. Biotite, and in some places, small grains of microcline associated with biotite took up the potassium released during crushing and recrystallization. Sphene has grown around the opaques. Scapolite, sericite, muscovite, and chlorite are rare in the crushed rocks.

The cataclasis and recrystallization of the anorthosites took place during the Grenville Orogeny. Mineral assemblages developed in these rocks at that time are indicative of amphibolite facies metamorphism.

The sequence of cataclastic deformation, metamorphism, and recrystallization as observed in the anorthosites is similarly represented in the rocks of the charnockitic suite which, in the undeformed state, are also orthopyroxene-bearing, coarse grained, porphyritic rocks. The main primary minerals include mesoperthitic orthoclase, plagioclase, orthopyroxene, clinopyroxene, hornblende, quartz, magnetite, zircon, and apatite. Cataclasis and recrystallization gave rise to simple mortar texture grading to well developed polygonal texture. In response to the amphibolite facies metamorphism garnet and sphene formed and the pyroxenes were transformed into olive green to brownish green hornblende and/or biotite. Extreme cataclasis and recrystallization reduced these rocks to medium to fine grained, equigranular, well foliated charnockitic gneiss. The gneiss lacks identifiable remnants of feldspar phenocrysts, and thus is difficult to distinguish from the granulite gneiss that developed near the anorthosite-mangerite masses.

General Remarks

The amphibolite facies is subdivided into the lower to middle amphibolite facies, the upper amphibolite facies, and the undivided amphibolite facies. The boundary between the middle and the upper amphibolite facies is a transition zone, marked in most cases by the appéarance of sillimanite with orthoclase in the presence of muscovite. Cordierite was observed in three places associated with sillimanite. In general, plagioclase composition in the amphibolite and the granulite facies gneisses lies near the oligoclase-andesine boundary. It is not possible therefore to distinguish between amphibolite facies and granulite facies rocks simply on the basis of plagioclase composition. The granulite facies rocks consist of two types of assemblages: one containing hornblende and/or biotite, and the other free of hydrous minerals. The great majority of the granulite facies rocks belong to the former.

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THE SIGNIFICANCE OF GARNET "ISOGRADS" IN GRANULITE FACIES ROCKS OF THE ADIRONDACKS

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Abstract

Garnet in Adirondack meta-igneous rocks is produced by complex reactions involving plagioclase and two or more mafic phases. Extrapolation of published experimental data yield rough estimates of 610° ± 30°C at 5.3 ± 1.5 kb and 810° ± 30°C at 6.8 ± 1.2 kb for the temperature and pressures at which these reactions occurred, in silica-saturated and undersaturated rocks respectively. The reactions are retrograde, occurring during cooling from igneous temperatures; therefore these temperature and pressure estimates do not represent maximum metamorphic conditions in the Adirondacks. Presence or absence of garnet in these rocks is controlled by kinetic factors rather than by thermodynamic equilibria. Garnet isograds mapped by Buddington and deWaard correspond to depths of emplacement, or maximum depths of burial, such that upon uplift and cooling the rocks entered the stability field of garnet at temperatures sufficiently high for the garnet-producing reactions to occur. Apparently anomalous occurrences of garnet on the low pressure side of these isograds are generally confined to unusually fine grained and/or intensely deformed rocks. A geothermal gradient of at least 30-35 C km of depth, and possibly higher, is inferred for the Adirondacks region during the Grenville event(s).

Résumé

Dans les roches méta-ignées des Adirondacks, les grenats sont produits par des réactions complexes, où sont en présence des plagioclases et au moins deux phases mafiques. Par extrapolation des données expérimentales publiées, on parvient à des estimations très approximatives, de 610° ± 30°C à 5.3 ± 1.5 kb et 810° ± 30°C à 6.8 ± 1.2 kb pour la température et les pressions auxquelles ont lieu ces réactions, dans les roches saturées en silice respectivement. Les réactions sont rétrogrades, et s'effectuent pendant l'abaissement de température au-dessous des températures de formation des roches ignées; par conséquent ces estimations de température et de pression ne représentent pas les conditions de métamorphisme maximum dans les Adirondacks. La présence ou l'absence de grenats dans ces roches dépend de facteurs cinétiques, plutôt que d'équilibres thermodynamiques. Les isogrades des grenats cartographiés par Buddington (1963) et deWaard (1965) correspondent aux profondeurs de mise en place ou bien aux profondeurs maxima d'enfouissement, où pendant leur soulèvement et refroidissement, les roches sont entrées dans le champ de stabilité des grenats, à des températures suffisamment élevées pour qu'aient lieu les réactions de formation des grenats. Apparemment, la formation anormale de grenats du côté de ces isogrades caractérisé par de faibles pressions est généralement limitée à des roches inhabituellement fines ou intensément déformées, ou les deux à la fois. On en déduit pour la région des Adirondacks, pendant le ou les événements de Grenville, un gradient géothermique d'au moins 30 à 35 C par kilomètre de profondeur, et peut-être même encore plus élevé.

INTRODUCTION

The presence or absence of garnet in meta-igneous rocks of the Adirondack area has frequently been used to estimate the conditions of metamorphic history of the area (Buddington 1963, 1965; deWaard, 1965, 1967a, 1967b, 1969, 1971b; Whitney and McLelland, 1973). Buddington (1963, 1965) proposed garnet isograds in the northwest Adirondacks (Fig. 1) representing the first appearance of garnet in biotite-quartz-plagioclase paragneisses, (olivine) metadolerite and metagabbro gneiss, and in syenite and quartz syenite orthogneisses respectively. He interpreted this sequence as a record of increasing temperature and load pressure in an eastward direction, i.e. as evidence that the Adirondack Highlands represent a more deeply eroded terrane than the northwest lowlands. Buddington (1965) emphasized the irregular occurrence of garnet in these rocks and suggested the importance of kinetic factors in determining whether or not garnet appears in a rock of given bulk composition under a particular set of P-T conditions. Intensity of deformation, grain size, and pressure and composition of a fluid phase are interrelated variables affecting the reaction rates; Buddington (1963) pointed out the common occurrence of garnet in sheared and recrystallized zones whereas unrecrystallized host rock of nearly identical composition contains no garnet. Buddington (1965, p. 77), noting that the role of temperature is also ambiguous, commented as follows:

> "The (garnet isograd in metagabbroic rocks) is primarily the consequence of an increase in temperature, but the effect of the temperature may be interpreted either as that requisite for the development of garnet as a member of an equilibrium assemblage or as that necessary to afford the rate of reaction for the development of garnet under conditions where it was already a potentially stable mineral."

An additional factor considered by Buddington (1965) is the role of bulk composition, in particular the Al_2O_3/CaO ratio in non-garnetiferous amphibolites as compared with that in garnetiferous amphibolites. deWaard (1965, 1967a) used an ACF plot of chemical analyses of garnetiferous and non-garnetiferous amphibolites, metagabbros and mafic granulites to demonstrate that rocks on the high-Al_2O_3 side of the plagioclase-hornblende field consistently contain garnet (assemblage plagioclase-garnet-hornblende ± orthopyroxene ± biotite) whereas those on the high CaO side do not (assemblage plagioclase-clinopyroxene-hornblende ± orthopyroxene). deWaard (1965, 1967a, 1967b) proposed a single garnet "isograd" based on the appearance of garnet in quartzofeldspathic rocks and

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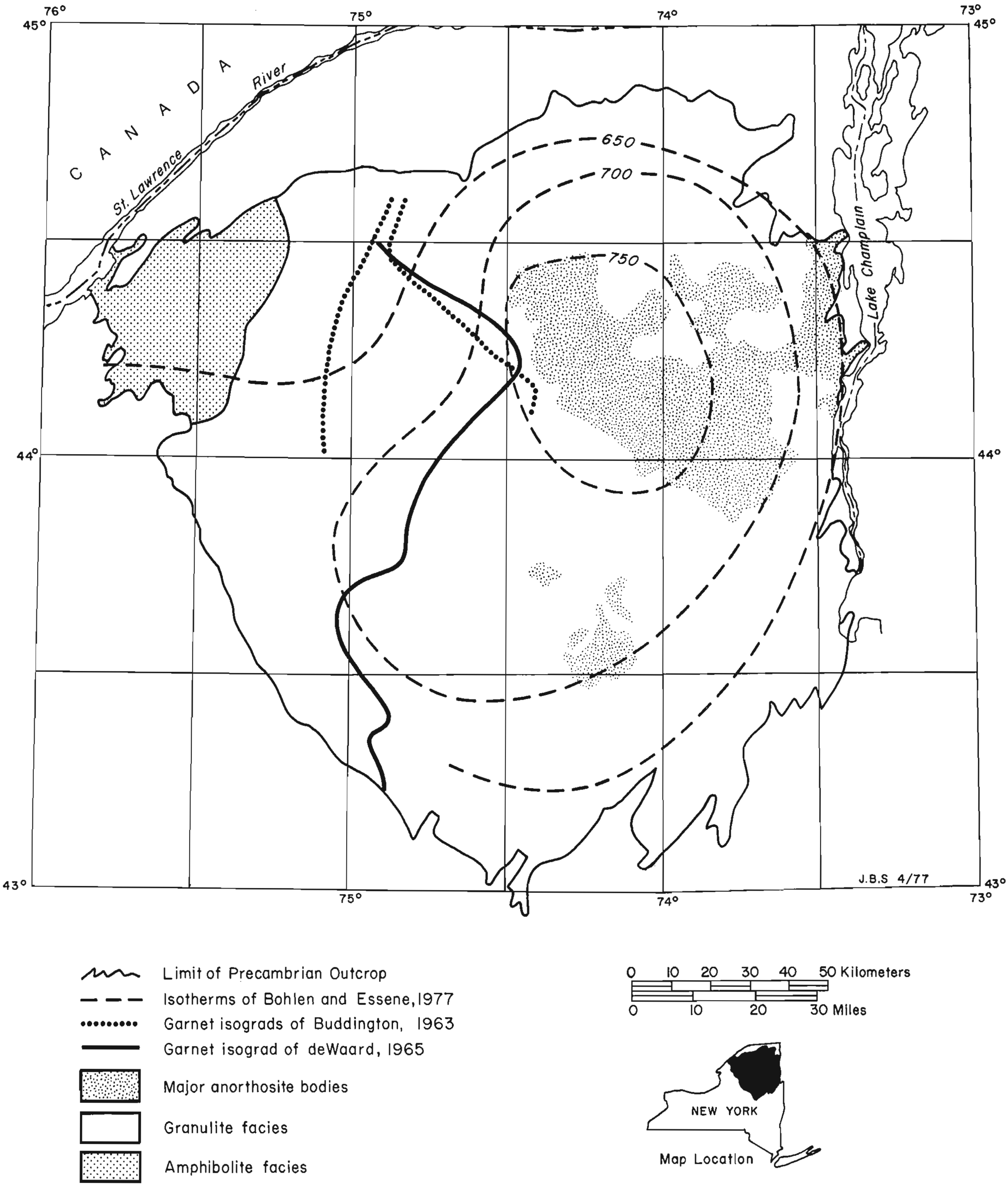

Figure 1. Outline map of the Adirondack region. Dotted lines are Buddington's (1963) second (to west) and third garnet isograds; solid line is garnet isograd proposed by deWaard (1965). Garnet is stable on east side of each isograd. Isotherm values are degrees Celsius.

suggested (deWaard, 1965) that this isograd represents the reaction:

$$6 \text{ orthopyroxene} + 2 \text{ anorthite} = \text{clinopyroxene} + \text{garnet} + 2 \text{ quartz} \qquad \text{(R.1)}$$

This isograd is also shown on Figure 1; as expected the northern portion is close to Buddington's third isograd, since the two are essentially equivalent. deWaard (1969, 1971a, 1971b) interpreted this boundary as resulting from the appearance of garnet during prograde metamorphism of increasing intensity in an eastward direction across the Adirondack Highlands.

MODE OF OCCURRENCE OF GARNET IN ADIRONDACK META-IGNEOUS ROCKS

A. Quartz-saturated Rocks

The quartz-saturated meta-igneous rocks of the Adirondacks include quartzofeldspathic gneisses containing plagioclase, alkali feldspar (commonly as mesoperthite), quartz, hornblende and clinopyroxene, with or without orthopyroxene, garnet, biotite, ilmenite and magnetite. Where these rocks contain hypersthene, they are commonly referred to as the mangerite-charnockite series. Similar rocks without hypersthene have usually been called "granite-gneiss" or more commonly "hornblende granite-gneiss". Some of these rocks, such as the Tupper, Stark and Diana complexes (Buddington and Leonard, 1962) are clearly igneous intrusives; others, commonly occurring as conformable units in the regional stratigraphy, are of more enigmatic origin, and may be metavolcanics (Turner, 1971). Most of the anorthositic rocks of the Adirondacks, as well as some of the associated gabbroic anorthosites, leuconorites and oxide-rich gabbros, contain normative and/or modal quartz in small amounts.

Garnet may occur in three distinct forms in these rocks:

1. as discrete grains or porphyroblasts either isolated or associated with schlieren of other mafic minerals in rocks with strong metamorphic fabric.

2. as rims or coronas around grains of ilmenite and/or magnetite in contact with plagioclase. The garnet in these rims appears to embay and replace both the plagioclase and the oxide phase. Ilmenite within such coronas may be partly replaced by one or more titanium-rich minerals (McLelland and Whitney, 1977).

3. as rims or coronas around orthopyroxene, orthopyroxene partially or wholly replaced by clinopyroxene, or clusters of ortho- and clinopyroxene grains. The garnet tends to be intergrown with a clear mineral of low birefringence, which may be either plagioclase (McLelland and Whitney, 1977) or quartz (deWaard, 1965; Davis, 1971). The garnet corona is commonly in contact with, and replaces, peripheral plagioclase. It is separated from internal pyroxenes by a zone of clear plagioclase and/or quartz.

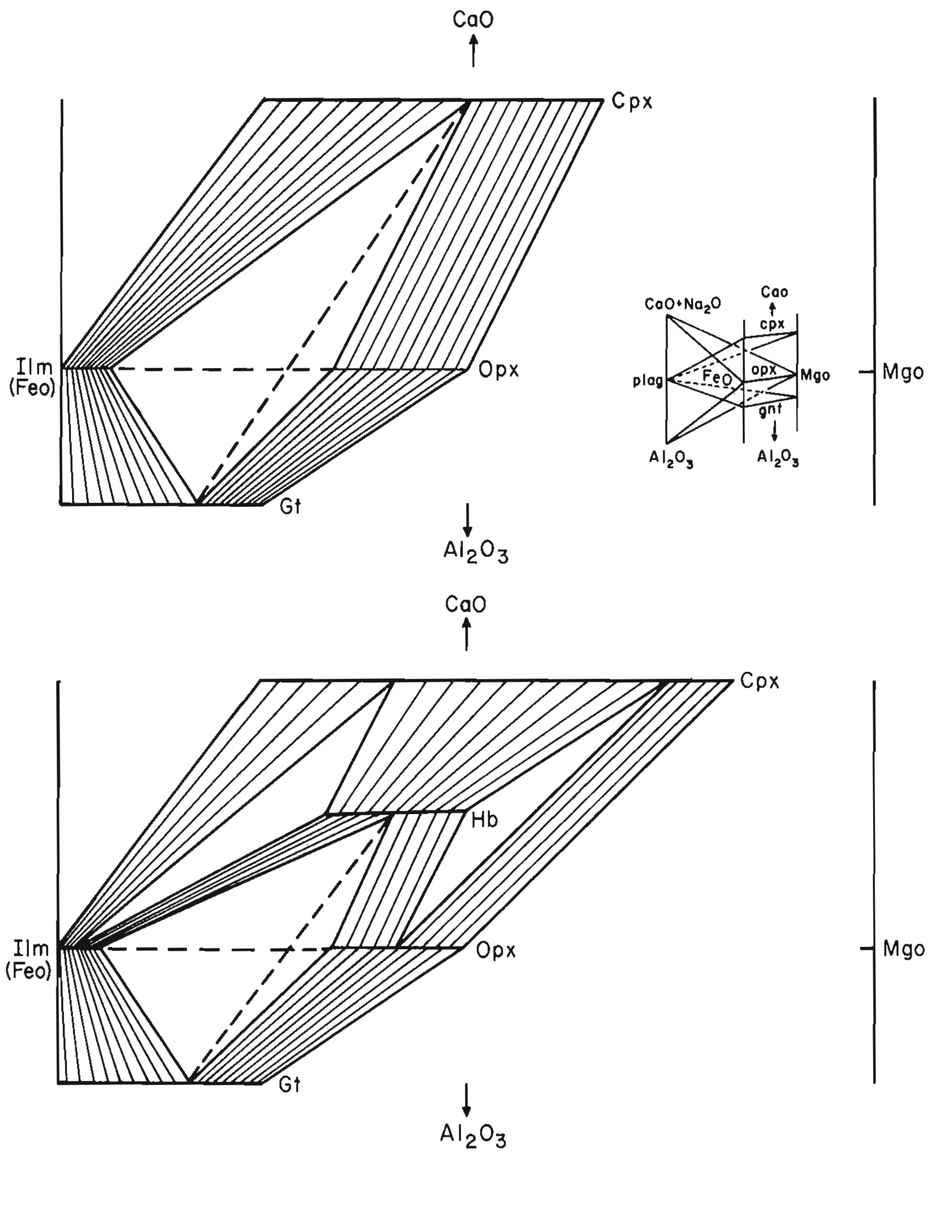

Figure 2.

Hypothetical ACFM projections for quartz-saturated rocks without hornblende (a) and with hornblende (b). Inset shows nature of projection.

The latter two textural forms may in some cases be combined with a single corona of garnet surrounding both pyroxene and the oxide phase(s). These garnets lack significant internal zoning, and within a thin section are of identical composition whether they occur around oxides or pyroxenes (McLelland and Whitney, 1977). Plagioclase near but external to garnet is more sodic than that in the interiors of large plagioclase grains or remote from the coronas.

The close similarity in composition between garnets in the two different types of coronas suggests a common origin, by a reaction involving both orthopyroxene and ilmenite and/or magnetite as well as plagioclase. One possible form for the reaction (McLelland and Whitney, 1977) in relatively "dry" rocks (without significant hornblende or biotite) is

$$\begin{array}{lll} \text{anorthite} & \text{orthopyroxene} & \\ 2\,CaAl_2Si_2O_8 + & (6\text{-}X)(Fe,Mg)SiO_3 + & XFeO \text{ (from ilmenite or magnetite)} \\ + \text{ quartz} = & \text{garnet} \qquad + & \text{clinopyroxene} \\ (X\text{-}2)SiO_2 & Ca(Fe,Mg)_5Al_4Si_6O_{24} & Ca(Fe,Mg)Si_2O_6 \end{array} \qquad (R.2)$$

Where X is a function of the distribution coefficients for Fe and Mg among the ferromagnesian phases. Reaction (2) is a complex reaction similar to the cyclic reactions described by Carmichael (1969), and can be written as the sum of several partial reactions occurring at separated locations as much as a few centimetres apart. These partial reactions are detailed by McLelland and Whitney (1977). Note that reaction (1) is a special case of reaction (2) for iron-free or oxide-free rocks. If magnetite is involved as the oxide phase, as suggested by Bohlen and Essene (1975), reaction (2) may involve the concurrent reduction of ferric iron and release of oxygen; alternatively the ferric iron may be taken up in the form of acmite component in the clinopyroxene or andradite component in the garnet. Most available evidence, however, points to ilmenite as the oxide mineral involved in reaction (2).

A reaction taking place in the idealized CFMAS system (CaO-FeO-MgO-Al_2O_3-SiO_2) can be shown by projection from plagioclase in the tetrahedron CaO-MgO-FeO-Al_2O_3 onto a plane parallel to the CaO-Al_2O_3 edge (Fig. 2). CaO and Na_2O are treated as isomorphous components; quartz is assumed to be present along with plagioclase. The effects of Fe_2O_3 and TiO_2 are neglected. Similar projections have been employed for the CFMAS system by deWaard (1971a) and Loomis (1975). Figure 2a shows schematically the positions of clinopyroxene, orthopyroxene and garnet in this projection. The horizontal tieline ilmenite-orthopyroxene shows the system on the high temperature-low pressure side of reaction (2); most or all Adirondack anorthositic rocks and quartzofeldspathic gneisses of igneous origin plot on the CaO-rich side of this tieline and hence contain no garnet as long as orthopyroxene and ilmenite coexist stably with plagioclase. The stable assemblages will be plagioclase-clinopyroxene-orthopyroxene and plagioclase-clinopyroxene-orthopyroxene-ilmenite. Upon increasing pressure or decreasing temperature the tieline clinopyroxene-garnet becomes stable, giving the assemblages plagioclase-clinopyroxene-orthopyroxene-garnet and plagioclase-clinopyroxene-garnet-ilmenite. This "dry" system, lacking hornblende and biotite, is realized to a first approximation in many anorthosites and anorthositic gabbros, and in a few of the charnockitic rocks. Equilibrium has clearly not been achieved in the coronitic textures, but a closer approach to equilibrium may be demonstrated by the fine grained, oxide-rich "Woolen Mill Gabbro" in the Elizabethtown Quadrangle, described by Buddington (1939, 1952). Texturally this rock is a granulite, comprising the assemblage oligoclase-clinopyroxene-garnet-ilmenite-magnetite with minor quartz. The relatively sodic plagioclase in this rock is consistent with the consumption of anorthite in reaction (2).

Increasing P_{H_2O} to the point where hornblende becomes a stable phase in there rocks may give the result shown in Figure 2b. Lacking analyses of hornblende coexisting with pyroxene and garnet from these rocks, the positions of the phases in the projection are only approximate. The topology as shown is based on the commonly observed sequence of $MgO/MgO{+}FeO$ ratios in mafic minerals in high grade metamorphic rocks, i.e. clinopyroxene > hornblende > orthopyroxene > garnet. Hornblende may now be produced in the garnet forming reaction in place of clinopyroxene. Garnet will not be stable in rocks having Al_2O_3/CaO ratios less than that in hornblende, in keeping with deWaard's (1965, 1967) observations. The frequently observed assemblage plagioclase-garnet-hornblende-clinopyroxene ± orthopyroxene probably represents rocks with relict igneous clinopyroxene; this clinopyroxene frequently exhibits hornblende reaction rims.

B. Undersaturated Rocks

The silica-undersaturated rocks of the Adirondacks include the olivine metagabbros, metadiabases and associated amphibolites. These occur for the most part as large, partly layered bodies on the eastern and southern periphery of the Marcy anorthosite massif, and as smaller bodies, dykes and conformable layers within metasediments, metavolcanics and quartzofeldspathic gneisses throughout the Adirondack region.

In the undersaturated rocks, as in the anorthositic and charnockitic rocks, garnet occurs in several texturally distinct forms:

– as discrete porphyroblasts, in places up to several centimetres and even several decimetres in diameter, in rocks with a distinct metamorphic fabric (amphibolites and mafic granulites).

– as complex coronas of radially oriented hypersthene around olivine (Whitney and McLelland, 1973) and surrounded by a thin band of clinopyroxene. The two pyroxene shells may be indistinct or replaced by a single shell of mixed clino- and orthopyroxene. A thin shell or "moat" of clear plagioclase commonly encloses the pyroxene shell(s), and in turn is

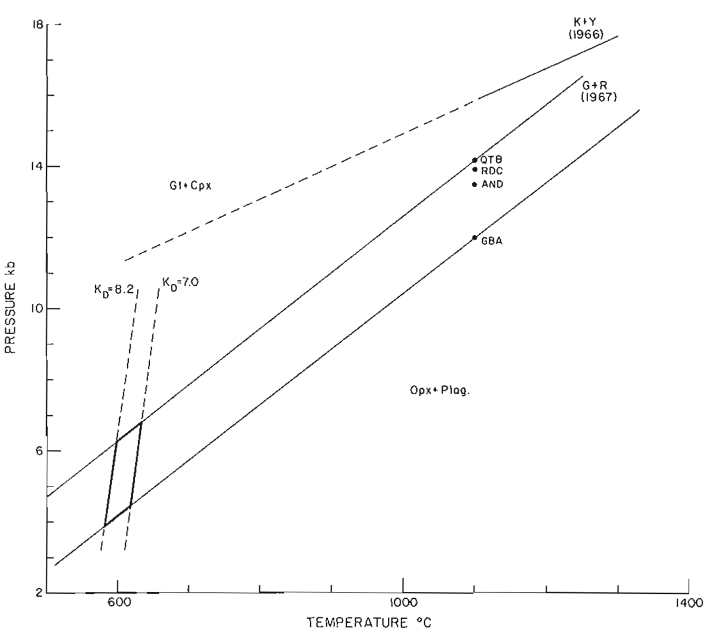

Figure 3. Pressure-temperature diagram for garnet-forming reactions in quartz-saturated rocks.

surrounded by a shell of garnet containing inclusions of clinopyroxene or, less commonly, green spinel and/or hornblende. The garnet shell replaces primary igneous cumulate plagioclase, which is clouded with minute green spinel crystals (Whitney, 1972).

– as coronas around oxide minerals, usually ilmenite. In places, garnet forms the outermost shell of the corona sequence: ilmenite and/or magnetite and/or green spinel, biotite, hornblende, garnet. This sequence is ubiquitous in relatively undeformed olivine metagabbros, but garnet, which replaces primary glagioclase, is of sporadic occurrence. The hornblende, where not rimmed by garnet, also replaces plagioclase, and commonly contains trains of small green spinel crystals inherited from the plagioclase (Whitney and McLelland, 1973). The garnet is nearly identical to that in the coronas around olivine, but may be very slightly richer in Ca and Fe, and poorer in Mg.

In the undersaturated rocks, the garnet in the olivine coronas is evidently the product of two consecutive reactions (Whitney and McLelland, 1973). In simplified form, these are:

$$\underset{\text{olivine}}{2\,(Mg,Fe)_2SiO_4} + \underset{\text{anorthite}}{CaAl_2Si_2O_8} = \underset{\text{orthopyroxene}}{2\,(Mg,Fe)SiO_3} + \underset{\text{clinopyroxene}}{Ca(Mg,Fe)Si_2O_6} + \underset{\text{spinel}}{(Mg,Fe)Al_2O_4} \qquad \text{(R.3a) and}$$

$$\underset{\text{orthopyroxene}}{4\,(Mg,Fe)SiO_3} + \underset{\text{anorthite}}{CaAl_2Si_2O_8} + \underset{\text{spinel}}{(Mg,Fe)Al_2O_4} = \underset{\text{garnet}}{Ca(Mg,Fe)_5Al_4Si_6O_{24}} \qquad \text{(R.3b)}$$

The products of reaction (3a) are preserved in places where, due to lower pressures (as in the western Adirondacks) and/or slower reaction rates, reaction (3b) has failed to occur, or occurred only to a limited extent. In these cases the coronas have a similar form, but the outer shell adjacent to primary plagioclase is composed of a complex intergrowth of clinopyroxene and green spinel rather than garnet. In the majority of undersaturated rocks from the Adirondack Highlands, however, garnet has replaced most or all of the clinopyroxene-spinel symplectite.

Reactions (3a) and (3b), like reaction (2) in saturated rocks, comprise several intermediate reactions (Whitney and McLelland, 1973). One result of this is the formation, during reaction (3a), of green spinel in two distinct textural forms, as symplectic intergrowths with clinopyroxene in the outer shell, and as microscopic inclusions producing intense clouding in the primary plagioclase. The former variety of spinel is largely exhausted during reaction (3b) whereas the latter persists and is augmented during growth of garnet.

Hornblende, nucleating around the opaque oxides, replaces spinel-bearing plagioclase, and is sporadically rimmed and replaced by garnet. This clearly points to the paragenetic sequence spinel → hornblende → garnet; hence the hornblende probably forms after the initiation of reaction (3a) and before the cessation of (3b). The hornblende-producing reaction has not been studied in detail, but the general form may be similar to:

$$\text{plagioclase + orthopyroxene + clinopyroxene + Fe oxides + spinel} + H_2O = \text{hornblende} \qquad \text{(R.4)}$$

Although textural relationships suggest the involvement of the oxides and/or hornblende in the formation of garnet, reaction (3b) has occurred in some rocks nearly devoid of either oxides or hornblende. Sack (1977) has demonstrated that magnetite is consumed during formation of garnet in plagioclase – magnetite cumulates in a layered metagabbro at the Split Rock Mine near Westport, N.Y.; here also hornblende appears to precede garnet in the paragenetic sequence.

PRESSURE-TEMPERATURE CONDITIONS OF GARNET-PRODUCING REACTIONS

Much experimental work has been done relevant to the occurrence of garnet in metabasic rocks because of considerable interest in the minerology of the upper mantle and in the pressure-temperature constraints on the gabbro-eclogite transition (Kushiro and Yoder, 1966; Green and Ringwood, 1967, 1972; Green and Hibberson, 1970; Irving and Green, 1970; Ito and Kennedy, 1971). Fewer studies have been made on the occurrence of garnet in quartzofeldspathic rocks; data for these rocks have been reported by Green and Lambert (1965). It is convenient for the purposes of this discussion to consider separately the silica-saturated and silica-undersaturated cases.

A. Silica-saturated Rocks

Kushiro and Yoder (1966) investigating reaction (1) using a starting mixture of anorthite plus enstatite, derived the line shown in Figure 3, that has a positive slope of approximately 9 bars*/°C. All their data were obtained at temperatures above 1100°C, and extrapolation was made to lower temperatures assuming linearity. This line marks the appearance of garnet with increasing pressure or decreasing temperature in an iron-free system. Note that reaction (1), for an iron-free system, is a special case of reaction (2). Addition of FeO, a component present in the pyroxenes and in one or more oxide phases, may be expected to stabilize garnet at lower pressures and/or higher temperatures. Addition of albite to plagioclase will probably have the opposite effect. In the anorthositic and charnockitic rocks of the Adirondack and Morin areas, garnet is more likely to form where the Mg/(Mg + Fe) ratio is less than about 0.45, and the plagioclase relatively calcic (Martignole and Schrijver, 1973). The effect of iron appears to be quantitatively more important than that of sodium; this is also evident from data on basic granulites and anorthosites from West Bengal reported by Manna and Sen (1974). Further confirmation comes from experimental data for actual rock compositions, summarized by Green and Ringwood (1972). The pressures at which garnet first appears when the temperature is 1100°C, is shown in Figure 3 for four quartz-saturated rocks: quartz tholeiite (QTB), gabbroic anorthosite (GBA), andesite and quartz diorite (AND) and rhyodacite (RDC). Pressures range from about 12 kb (gabbroic anorthosite) to a little more than 14 kb (quartz tholeiite); all are significantly lower than the 15.8 kb reported by Kushiro and Yoder (1966) for the Fe- and Na-free system. The actual garnet-producing reaction in these experiments is not known, but the disappearance of iron oxides from the quartz tholeiite sample coincident with the appearance of garnet (Green and Ringwood, 1967, Fig. 5) suggests a reaction similar to (2).

As in the case of the artificial system, there is considerable uncertainty in the slope of the line dividing the garnetiferous and non-garnetiferous fields, and the location of the boundary below 1000°C must be derived by linear extrapolation. Green and Ringwood (1967) used the appearance of garnet in their quartz tholeiite "B" composition as the lower limit of the garnet granulite field. A slope for this boundary of roughly 15.7 bars/°C can be inferred from their data (Fig. 3) which, together with a parallel line drawn through the point for gabbroic anorthosite (GBA) defines a band in the P-T plane within which garnet probably first appears in rocks of composition comparable to Adirondack anorthositic and charnockitic rocks.

Raheim and Green (1974) have reported experimental data on distribution coefficients for iron between coexisting garnet and clinopyroxene in eclogites; i.e.

$$K_{D_{Fe}}^{Gt\text{-}Cpx} = \left(\frac{FeO}{MgO}\right)_{Gt} \Big/ \left(\frac{FeO}{MgO}\right)_{Cpx}$$

* 1 kb = 10^5 kPa

Because these data were obtained at pressures at or above 20 kb, well within the eclogite field, considerable extrapolation is necessary to apply them to rocks formed at pressures within the garnet granulite field. The intersection of lines of equal K_D with the lower pressure limit of garnet stability nonetheless provides a rough method for estimating the pressures and temperatures at which the garnet-producing reaction occurred, assuming that the garnet and clinopyroxene have not subsequently re-equilibrated. Data on coexisting garnet and clinopyroxene from saturated rocks of the Adirondack Highlands from Buddington (1952) and McLelland and Whitney (1977) show K_D values ranging from 6.7 to 8.2. The intersection of these two bands with the curves for the garnet-forming reaction forms a parallelogram (Fig. 3) indicating reaction temperatures on the order of 580-640°C at pressures of 3.8-6.8 kb. It should be emphasized that this argument relates only to reactions in the dry system illustrated in Figure 2a. Reactions involving hornblende, which may be more common in Adirondack rocks, have not been investigated experimentally.

B. Undersaturated Rocks

The first detailed experimental work on garnet-forming reactions in simplified analogs of undersaturated basic rocks was reported by Kushiro and Yoder (1966) who studied reactions (3a) and (3b) in the end-member system forsterite-anorthite; the resulting curves are shown in Figure 4. The position of the curve for reaction (3a) will clearly be influenced in natural systems both by the presence of FeO in olivine and Na_2O in plagioclase. The relative magnitudes of these effects were tested by Emslie (1971) using as reactants olivine (Fo_{70}) and plagioclase (An_{73}). The olivine metagabbros of the Adirondacks contain normative olivine of comparable composition and normative plagioclase ranging from An_{55} to An_{65}. Emslie's experiments showed that the effect of sodium on this reaction outweighs that of iron; the upper stability limit for the plagioclase-olivine assemblage is displaced toward higher pressures by roughly 2 kb. All experimental data were obtained at temperatures in excess of 1000°C and linear extrapolation was made to lower temperatures (Fig. 4). Obata (1976) calculated the equilibrium curve for reaction (3a) in the forsterite-anorthite system from thermodynamic data and found that it bends toward the temperature axis at lower temperatures. This calculation was done only for the end-member system, and the possible effects of FeO and Na_2O on the shape of the curve are uncertain. A parallel curve in Figure 4 extends Emslie's data to lower temperatures. The corresponding reaction in nature is multivariant so that this curve represents the first appearance of the spinel- two-pyroxene assemblage with rising pressure and/or falling temperature. A Schreinemakers analysis of the CMASH (CaO-MgO-Al_2O_3-SiO_2-H_2O) system by Frost (1976) indicates that at lower pressures in the presence of a vapour phase, plagioclase and olivine may react during cooling to produce the assemblage amphibole-clinopyroxene-spinel in place of the assemblage orthopyroxene-clinopyroxene-spinel which, at higher pressures, appears first.

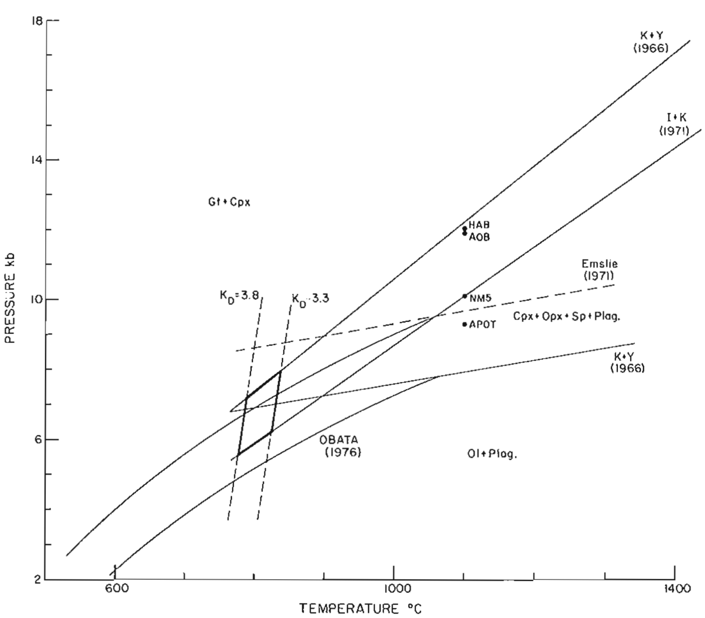

Figure 4. Pressure-temperature diagram for garnet-forming reactions in undersaturated rocks.

The P-T conditions for the garnet-producing reaction (3b) in undersaturated rocks as determined by Kushiro and Yoder (1966) in the forsterite-anorthite system are shown by the upper line in Figure 4. Experimental data for natural rocks, summarized by Green and Ringwood (1972), include a high alumina basalt (olivine normative), and alkali olivine basalt, and two olivine tholeiites. One of the olivine tholeiites, sample NM5, was studied by Ito and Kennedy (1971) who calculated a slope of 14 bars/°C for the garnet-in reaction curve. The alkali olivine basalt (AOB in Fig. 4) matches the Adirondack olivine metagabbros most closely in composition (unpublished data of the author). The alkali-poor olivine tholeiite (APOT) is least similar to the Adirondack rocks; the normative plagioclase of An_{83} (Green and Ringwood, 1967) is more calcic by about 25 per cent An and therefore the garnet is probably stable at a lower pressure than garnet in the Adirondack gabbro. The band on the P-T plane in Figure 4 between the curves of Kushiro and Yoder (1966) and that of Ito and Kennedy (1971) for sample NM5 thus defines the most probable conditions for the first appearance of garnet in the Adirondack olivine gabbros. Distribution coefficients for iron between garnet and clinopyroxene in these rocks range from 3.3 and 3.8 (Whitney and McLelland, 1973); P-T lines for these values extrapolated from the experimental data of Raheim and Green (1974) intersect the garnet-in curves at a high angle. The area of intersection (Fig. 4) suggests that the garnet reaction occurred at temperatures between 780 and 840°C and pressures between 5.6 and 8.0 kb. If reaction (3b) occurred during cooling, as is probable, the textural evidence that (3a) occurred first suggests that pressures in the higher part of this range are more likely.

Conditions of garnet formation in Adirondack meta-igneous rocks are summarized in Table 1.

Table 1*

	Quartz saturated (anorthosite, anorthositic gabbro, charnockite)	Undersaturated (olivine metagabbro)
T	610 ± 30°C	810 ± 30°C
P	5.3 ± 1.5 kb	6.8 ± 1.2 kb

It should be re-emphasized that the data of Table 1 are dependent upon extrapolation of experimental data obtained in most cases at much higher temperatures and pressures.

Since the garnet-producing reactions are almost certainly retrograde (see below) these pressures and temperatures (Table 1) do not represent the maximum attained during the metamorphism. Other geothermometers and geobarometers have yielded little precise information on this subject.

* Recalculation of the T, P values in Table 1 utilizing (a) new data, (b) corrections for Fe''' using the method of Ryburn et al. (1975), and (c) a smaller correction for pressure dependence of $K^{Gt-Cpx}_{D_{Fe}}$ as suggested by Wood (1975), leads to revised values of 715 ± 25°C, 7.1 ± 1.4 kb for the saturated case. In the undersaturated case, the various corrections nearly cancel, and the net differences in the estimated T and P are negligible.

The pelitic and semipelitic rocks present in the metasedimentary sequence in the highlands contain sillimanite as the only Al_2SiO_5 phase, with the exception of a single reported occurrence of kyanite in the Blue Mountain Lake Quadrangle (Ennis Geraghty, pers. comm.). The significance of kyanite with respect to metamorphic grade at this locality is uncertain because the mineral coexists with sillimanite in ambiguous textural relationship. The common presence of wollastonite in calc-silicate rocks, particularly in close proximity to anorthosite bodies, and the single reported occurrence of monticellite in the Mt. Marcy Quadrangle (Baillieul, 1976), are likewise of little use in fixing pressure and temperature of metamorphism due to the complex interplay of temperature and partial pressures of H_2O and CO_2 in calc-silicate equilibria. Valley and Essene (1977) have shown that P_{CO_2} during metamorphism of the carbonates and calc-silicates of the Adirondacks may have been quite low, permitting formation of wollastonite and monticellite at relative low temperatures.

The occurrence of high-iron orthopyroxene (Fs_{95}) in quartz mangerite (Jaffe et al., 1975) on Pitchoff Mountain in the Mt. Marcy Quadrangle has been used by Jaffe et al. (1977) to argue for final equilibration at 600-800°C and 7-11 kb, based on the data of Smith (1971). Apparent stable coexistence of quartz and fayalite in an iron-rich granitic rock in the Cranberry Lake and Ausable Forks quadrangles (Buddington, 1939; Buddington and Leonard, 1962) indicates either lower regional pressures or later, shallower intrusion of this rock, which is largely undeformed in contrast to the gneisses on Pitchoff Mountain. Jaffe et al. (1977) also studied pyroxene exsolution phenomena in anorthosites and related metagabbros of the Mt. Marcy Quadrangle, which led to estimates of 900-1100°C and 7-11 kb for temperatures and pressures in a presumed gabbroic anorthosite magma at the time of crystallization, with a later re-equilibration at 500-700°C.

A more stringent constraint on pressure is provided by reaction (3a) in the olivine metagabbro and metatroctolite. Crystallization of these rocks probably took place between 1100 and 1200°C. Emslie's (1971) experimental data and clear textural evidence that the spinel and much of the pyroxene is of metamorphic rather than igneous origin place an upper limit of 9-10 kb on the pressure at the time of emplacement.

Bohlen and Essene (1977) determined a series of isotherms in the Adirondacks by means of microprobe analyses of coexisting magnetite and ilmenite, and coexisting feldspars, using the experimental data of Buddington and Lindsley (1964) for the oxides and that of Stormer (1975) for the feldspars. The isotherms show a roughly concentric pattern around the Adirondack Highlands (Fig. 1) with maximum temperatures on the order of 790 ± 50°C between Saranac Lake and Tupper Lake. Most of the rocks used in this work were quartzofeldspathic orthogneisses (charnockites and hornblende-bearing granitic gneisses). Such rocks in the Adirondacks are commonly migmatitic, and consequently the measured temperatures may represent, not metamorphic maxima, but the solidus in a relatively dry granitic or quartz dioritic system, analogous to the rocks studied experimentally by Brown and Fyfe (1970). The regional trends shown by the data may reflect variations in P_S, P_{H_2O}, bulk composition, or a combination of these factors. It is unlikely that they are simple isotherms. They do, however, give a minimum temperature for the last pervasive re-equilibration of these rocks.

Maximum pressures and temperatures recorded in the anorthositic and gabbroic rock are in the order of 900-1200°C and 7-10 kb. Temperatures in the quartzofeldspathic rocks in the central part of the highlands reached a minimum of 800°C. In both cases the temperatures, and probably the pressures as well, were considerably in excess of those at which the garnet-producing reactions appear to have occurred, and well outside the garnet stability field.

RETROGRADE NATURE OF THE GARNET REACTIONS

The garnet-producing reactions in anorthositic, charnockitic, and gabbroic rocks of the Adirondack Highlands are almost certainly retrograde. The data presented have shown that the reactions occurred at temperatures well below metamorphic maxima and there is also clear textural evidence that garnet has replaced earlier minerals, in particular, plagioclase. It is thus apparent from field and experimental data (Figs. 3, 4) that garnet may form from plagioclase-orthopyroxene assemblages simply from a fall in temperature, a conclusion reached by workers in other areas (Griffin, 1971; Martignole and Schrijver, 1971; Manna and Sen, 1974). Crystallization of garnet in response to a rise in temperature would, by contrast, involve a complex sequence of events including intrusion, cooling and uplift of the igneous rocks without reaction, followed by prograde metamorphism on a steep geothermal gradient, followed in turn by an abrupt increase in pressure due perhaps to overriding by large nappes or thrust sheets. While large nappe structures are known in the Adirondacks (deWaard, 1962; McLelland, 1977), they appear to be early in the tectonic history of the area and to predate, in part at least, the intrusion of the major anorthosite and gabbro bodies. If such a prograde process did occur, and the rocks entered the garnet field with rising temperature and pressure, the disequilibrium exhibited by the corona textures would not be expected. A prograde model involving synkinematic metamorphism does not explain the undeformed nature of the corona structures. The simpler model involving emplacement of the anorthosites and gabbros late in the tectonic history, accompanied by partial melting and re-equilibration of the quartzofeldspathic rocks, and followed by formation of garnet upon cooling, is therefore preferred. Late phases of deformation which accompanied this cooling, producing gneissic foliation in parts of the anorthosite and in the quartzofeldspathic rocks, had substantially ceased by the time the rocks entered the garnet field.

SIGNIFICANCE OF THE GARNET ISOGRADS

If the retrograde origin of the garnet is valid, the garnet "isograds" proposed by Buddington (1963) and deWaard (1965) cannot be isograds in the classical sense. It is proposed that these surfaces are kinetically controlled and are related to the depth of emplacement of the original igneous rock or, in the case of those quartzofeldspathic gneisses which may be of metavolcanic origin, the maximum depth of burial. This hypothesis is illustrated schematically in Figure 5. At the time of the last major igneous event a relatively steep geothermal gradient prevailed in the middle and upper portions of the crust, due to the large volumes of magma intruded. This gradient is represented by curve AO. As magmatic activity waned, the regional gradient gradually changed toward that represented by BO. The cooling path followed by rocks emplaced at depth D_1 is represented by line CE, which intercepts the garnet stability field at a temperature T_1. The extent to which reaction takes place in a given bulk composition will depend upon rock texture and the presence or absence of a fluid phase as well as upon the temperature. Similar rocks emplaced at a higher crustal level, D_2, will cool along FG, and intersect the garnet field at a lower temperature, T_2. Due to this lower temperature, the garnet-producing reaction will occur at a slower rate. At some temperature $T_X < T_2$, corresponding to an initial depth of emplacement $D_X < D_2$, the reaction will not occur within a finite interval of time and garnet will not form. The garnet

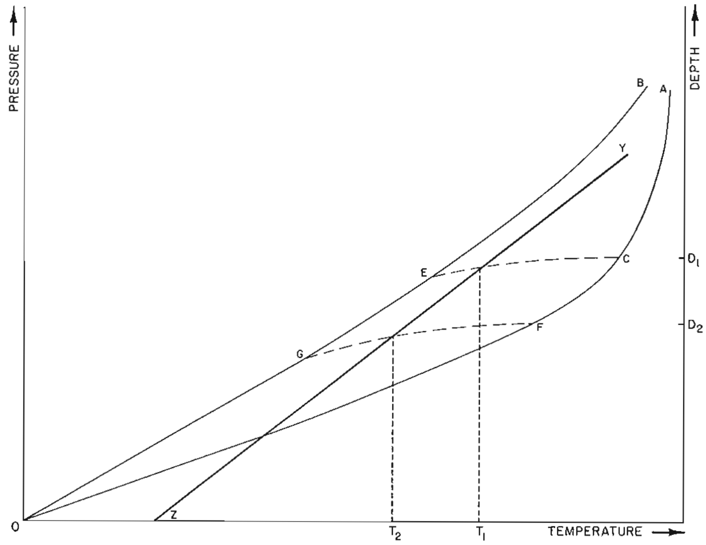

Figure 5. Schematic PT diagram illustrating hypothetical cooling paths (CE and FG) for Adirondack granulite facies rocks. Heavy line YZ represents high temperature, low pressure stability limit of garnet.

"isograd" as mapped will therefore represent this critical depth of emplacement. Variations in texture, activity of water, and bulk composition will broaden the kinetic isograd into a zone. The finer grained nature of cataclastically deformed rocks and the resulting increase in reaction rates probably explains Buddington's (1963) observation of garnet in shear zones in the northwest Adirondacks on the "wrong" side of the isograds.

Because mapped lines or zones representing the appearance of garnet in a particular rock type in granulite facies areas such as the Adirondacks are, by this hypothesis, not true isograds, they cannot be interpreted in precisely the same manner. It is possible, however, to view them as indicators of relative depths of erosion. For example, the garnet isograd mapped by deWaard (1965) in quartzofeldspathic gneisses separating his hornblende-orthopyroxene-plagioclase subfacies to the west from the hornblende-garnet-clinopyroxene subfacies to the east, corresponds to reaction (2) or to a similar reaction involving hornblende. In the more deeply eroded areas to the east, temperatures were sufficiently high for the reaction to occur. West of the isograd, the cooling curve intercepted the garnet stability field at temperatures too low for garnet to form in a finite time, except in unusually fine grained rocks. The position of the second isograd of Buddington (1965) (for metagabbroic rocks) to the west of the third isograd (for syenitic and quartz syenitic rocks) is consistent with the higher temperature of origin for garnet in undersaturated rocks as inferred from experimental data.

CONSTRAINTS ON THE GEOTHERMAL GRADIENT DURING GRENVILLE TIME IN THE ADIRONDACKS

If the garnet was produced during regional cooling, the temperature and pressure estimates in Table 1 indicate a geothermal gradient in the upper 20-25 km of the crust in the Adirondacks at that time of 30-35°C/km. This is a minimum estimate since the garnet was formed after regional cooling had commenced. It is compatible with the absence of retrograde kyanite in the metapelites of the Adirondacks, since with a gradient of 30°C/km or greater, the rocks would enter the kyanite field at temperatures under 600°C, which are probably too low for the sluggish inversion from sillimanite to take place. Further confirmation of a steep gradient was provided by the work of Stoddard (1976) on rocks near Colton in the northwest Adirondack lowlands, not far from the highlands boundary. On the basis of several criteria, he estimated peak metamorphic conditions of 735 ± 25°C; 4.5 ± 1.0 kb, necessitating a gradient in excess of 40°C/km in that area.

If the present crustal thickness is 35-40 km (Katz, 1955) and the presently exposed surface in the Adirondack Highlands corresponds to 8 kb (roughly 25 km) depth during Grenville time, the total crustal thickness was on the order of 60-65 km at that time. A gradient of 30-35°C/km in the upper portion of the crust, even allowing for considerable curvature toward the pressure axis in the lower crust due to depletion in radioactive elements, would be sufficient to allow for extensive partial melting and production of gabbroic anorthosite liquids near the base of a 60 km crust, consistent with the model of Simmons and Hanson (1977) for the origin of the Adirondack anorthosite.

CONCLUSIONS

1. Formation of garnet in the meta-igneous rocks of the Adirondack Highlands occurred by reactions between plagioclase, mafic silicates, and oxides, as described by Whitney and McLelland (1973) for olivine metagabbros and McLelland and Whitney (1977) for anorthositic and charnockitic rocks.

2. As first noted by Buddington (1965) presence or absence of garnet in these rocks depends largely on kinetic considerations. The garnet-producing reactions took place under retrograde (cooling) conditions; the cooling curves in all cases intercept the garnet stability field, but only at depths greater than a certain critical value do the rocks enter the garnet field at temperatures sufficiently high for reaction to occur in a finite time. This critical depth is influenced by rock texture and by the presence or absence of a fluid phase.

3. Garnet isograds in granulite facies terranes, such as those mapped by Buddington (1963) and deWaard (1965) in the Adirondacks, represent the critical depth of emplacement at which the garnet-producing reaction becomes kinetically possible.

4. Rough estimates of temperature and pressure for the formation of garnet in the Adirondacks , based on extrapolated experimental data, suggest a minimum geothermal gradient on the order of 30-35°C/km during the Grenville metamorphism.

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