

**LUNAR AND PLANETARY INSTITUTE    3303 NASA ROAD 1    HOUSTON, TEXAS 77058**

WORKSHOP ON  
A CROSS SECTION OF ARCHEAN CRUST

Edited by:  
L. D. Ashwal  
and  
K. D. Card

Field Guide and Abstracts Presented to the  
1983 Archean Geochemistry—  
Early Crustal Genesis Field Workshop

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## **Organizing Committee**

L. D. Ashwal  
Lunar and Planetary Institute

K. Burke  
SUNY at Albany and Lunar and Planetary Institute

K. D. Card  
Geological Survey of Canada

A. M. Goodwin  
University of Toronto

G. N. Hanson  
SUNY at Stony Brook

W. C. Phinney  
NASA Johnson Space Center

J. L. Wooden  
Lunar and Planetary Institute

## **Field Trip Leaders**

J. A. Percival  
Geological Survey of Canada, Ottawa

K. D. Card  
Geological Survey of Canada, Ottawa

L. S. Jensen  
Ontario Geological Survey, Toronto

L. E. Luhta  
Ministry of Natural Resources, Timmins

R. P. Sage  
Ontario Geological Survey, Toronto

---

Cover: *Deformed xenolithic tonalite gneiss metamorphosed to upper amphibolite facies at Missinaibi Lake, about 20 km east of the Michipicoten greenstone belt. Mafic xenoliths may have been derived from overlying metavolcanic units. Knife for scale.*



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Seated (left to right), Jimmy F. Wilson, Ullrich Rast, Ken Collerson, Elaine R. Padovani, Frank R. Luther, Lewis D. Ashwal, Greg M. Stott, Ken Card, Denis M. Shaw, Steven B. Shirey. Kneeling, Craig Schiffries, Peter Salpas, Kent C. Condie, John A. Percival, Paul A. Mueller, Gary P. Beakhouse, John W. Valley, David M. Fountain. Standing (left to right), Richard Ernst, Xianglin Qian, Norman M. Evensen, Gilbert N. Hanson, Larry A. Haskin, Werner Weber, Richard Stern, Roberta L. Rudnick, Robert A. Frith, Paul Sylvester, Marilyn Truscott, Robert F. Folinsbee, George R. Tilton, Cornelis Klein, Tomas Feininger, George E. McGill, Klaus J. Schulz, William C. Phinney, J. D. Macdougall, Lincoln S. Hollister, Richard Carlson, Darrell Henry, Jim Papike, Robert F. Dymek, S. Ross Taylor. Not shown, LeBecca Turner (photographer), Larry Jensen, Lorne E. Luhta, Ron P. Sage, Scott Young.

# Introduction

An issue of major importance in crustal genesis is the relationship between Archean low- and high-grade terrains. Some workers view the high-grade gneiss regions as deeply eroded equivalents of the lower grade granite-greenstone terrains, whereas others regard the contrasting regions as the products of differing tectonic environments.

An ideal opportunity to examine these relationships exists in the central Superior Province of the Canadian Shield. Here, a 120-km-wide transition from subgreenschist facies rocks of the Michipicoten greenstone belt to granulite facies rocks of the Kapuskasing structural zone represents an oblique cross section through some 20 km of crust, uplifted along a northwest-dipping thrust fault. This area is unique because of its relative accessibility, because the rocks involved are about 2700 m.y. old, and because it is apparently uninterrupted by major faults, thus providing a continuous profile into lower continental crust of Archean age.

Recognizing these attributes, a field workshop was organized with major funding from NASA and the sponsorship of other U.S., Canadian, and international organizations. In keeping with the overall goals of the Early Crustal Genesis Program, the workshop was designed to foster interaction between the planetary and Archean scientific communities. Activities included a day of oral presentations and attendant discussions in Ottawa, and a six-day field trip to the Wawa-Chapleau-Timmins region of northeastern Ontario. This volume contains extended abstracts from the speakers at the technical session, additional contributed abstracts, and a guidebook to the field trip area. Also included is a summary of group discussions on the prospects for future activities, publications, research in the field trip area either proposed or in progress, and a compilation of suggestions and ideas for future research.

The success of this workshop can be measured not only by the highly inspiring technical presentations and field trips, but also by the new research that was spawned, some involving multidisciplinary collaborative efforts. Thus we expect the research stimulated here to bear fruit for some time to come.

# Program

**Early Crustal Genesis Field Workshop**  
**August 10, 1983**  
**Alice Wilson Hall, Geological Survey of Canada**  
**Ottawa, Ontario**

**Conveners: L. D. Ashwal and K. D. Card**

**8:00 AM     Registration**

**9:00 AM     Morning Session**

John Fyles, Geological Survey of Canada  
*Introduction and Welcoming Remarks*

J. A. Percival and K. D. Card, Geological Survey of Canada  
*Field Trip Orientation: The Archean Crust in the Wawa-Chapleau-Timmins Region*

**9:30 AM     Presentations**

Wilson J. F.  
*The Greenstone Belts of Zimbabwe*

Condie K. C.     Allen P.  
*The Transition from an Archean Granite-Greenstone Terrane into a Charnockite Terrane in Southern India*

Fountain D. M.  
*Geophysical Consequences of Phanerozoic and Archean Crustal Evolution: Evidence from Crustal Cross-Sections*

Weber W.  
*The Pikwitonei Granulite Domain: A Lower Crustal Level Along the Churchill-Superior Boundary in Central Manitoba*

Ashwal L. D.     Morgan P.     Leslie W. W.  
*Thermal Constraints on High Pressure Granulite Metamorphism of Supracrustal Rocks*

Rudnick R. L.     Ashwal L. D.     Henry D. J.  
*Metamorphic Fluids and Erosion History of a Portion of the Kapuskasing Structural Zone, Ontario as Deduced from Fluid Inclusions*

Hanson G. N.  
*Use of Olivine and Plagioclase Saturation Surfaces for the Petrogenetic Modeling of Recrystallized Basic Plutonic Systems*

**12:00 Noon     Lunch Break**

**1:30 PM     Afternoon Session**

Taylor S. R.  
*Continental Crustal Composition and Lower Crustal Models*

Dymek R. F. Boak J. L. Gromet L. P.

*Average Sedimentary Rock Rare Earth Element Patterns and Crustal Evolution: Some Observations and Implications from the 3800 Ma Isua Supracrustal Belt, West Greenland*

Rast U.

*A Multi-Element Study of Isua Iron-formation, W-Greenland*

McGill G. E.

*Venus Topography: Clue to Hot-Lithosphere Tectonics?*

Tilton G.R.

*Archean Crust-Mantle Geochemical Differentiation*

Carlson R. W. Hunter D. R. Barker F.

*Sm-Nd Isotopic Systematics of the Ancient Gneiss Complex, Southern Africa*

Shirey S. B. Hanson G. N.

*Granitic Rocks and Metasediments in Archean Crust, Rainy Lake Area, Ontario: Nd Isotope Evidence for Mantle-like Sm/Nd Sources*

Maccougall J. D. Gopalan K. Lugmair G. W. Roy A. B.

*An Ancient Depleted Mantle Source for Archean Crust in Rajasthan, India*

Collerson K. D.

*Ion Microprobe Zircon Geochronology of the Uivak Gneisses: Implications for the Evolution of Early Terrestrial Crust in the North Atlantic Craton*

**4:30 PM Adjourn**

# Summary of Discussions

Compiled by W. C. Phinney

At the conclusion of the last formal day of the field workshop in Timmins, Ontario on August 15 the participants convened for a discussion of the Early Crustal Genesis Program, including suggested activities for the future, nature of publications associated with the program, critique of the workshop, research ideas that were suggested during the meeting, and planned research activities in the field workshop area by the participants. Following the discussion the participants were requested to write down their suggestions for research ideas, their planned research in the field trip area, and any other comments that they felt to be appropriate.

Part A is a compilation of the suggestions for research ideas from the participants, Part B is a compilation of the studies that are either underway or planned by the participants, and Part C is a compilation of points made during the discussion.

## Part A. Suggestions for Research Ideas

### 1. Basic Data Needed

a. A detailed structural study of the Wawa greenstone belt is necessary to truly define the stratigraphic sequences of volcanic flows and various sediments, including iron formations. This would lead to better constraints for tectonic models, pressure determinations, chemical variations between and within volcanic centers, and the relationship between sources of sediments and volcanic or other igneous centers, and would provide a sound basis for tracing units into higher grade zones where lower crustal processes are being studied.

b. In order to characterize the protolith of supracrustal rocks in high-grade terrains, there should be an emphasis placed on finding areas where stratigraphic units can be traced across low-grade to high-grade metamorphic boundaries (for example, felsic volcanic rocks within stratigraphic sequences). This is extremely crucial to many geochemical and petrologic studies involving metamorphic reactions, fluid migrations, diffusion of components, isotopic reequilibrations, nature of partial melting, and several others. More detailed mapping is needed throughout the area with these objectives in mind.

c. In view of overturning in the Wawa greenstone belt and the flat lying domal structures in the middle- to high-grade zones, a thorough search for nappes and recumbent folds seems in order. What is the extent of doming by truly diapiric tonalites? Are the alleged domes really domes or are they a result of cross folding? This information is crucial to the development of tectonic models.

d. In the high-grade terrains, can some of the mafic gneisses be demonstrated as residues of the presumed partial melting event that formed the ubiquitous tonalites? Such areas are critical for the proper collections of samples that are used for detailed determinations of lower crustal processes.

e. Using the boulders in the conglomerates and inclusions in the diatremes, a search for traces of an ancient crust is possible in this area. For example, 3200 m.y. zircons from tonalitic gneiss boulders in a 2600 m.y. diatreme were discovered in the Yellowknife area of Canada. Thorough collections for this purpose from such occurrences in the field trip are advisable if a history of crustal evolution is to be developed.

f. On a general scale, comparison of many aspects of Archean and Tertiary high-grade lower crustal associations should be made in order to determine which rock types, structures, metamorphic processes, and tectonic processes are truly unique to the Archean.

g. Determine the variations in thickness, trend, and dip of mafic dikes of various ages across large regions of the Superior province to yield potential information on the depths of exposure at various times and locations during and after the Archean.



h. A crustal cross section of the type in the field trip area requires full geophysical characterization on a regional scale: seismic studies (both refraction and reflection) that might allow tracing of specific units from the surface to depth or from one surface exposure to another, electrical conductivity studies, magnetic surveys, gravity surveys on a variety of scales, and heat flow measurements.

i. Geophysical characterization of the range of individual rock types from the cross section are necessary for interpretation of the regional data. This includes measurements of sonic velocities at various P and T conditions, thermal conductivities, heat production, magnetic properties, densities, and electrical conductivities.

j. It is necessary to characterize the various rock types and their relationships with each other throughout all of the metamorphic zones. This includes mineral assemblages and textures.

## **2. Processes in the Lower Crust**

a. Determine the role of fluids during granulite facies metamorphism. Both the changes in composition of fluids and the origin of the fluids require study. Fluid inclusions in rocks sampled across the facies changes may provide information on changes in composition. Also, changes in the nature of fluid inclusions across pyroxene granulite xenoliths rimmed with amphibolite in tonalite should provide insight into the types of fluids associated with the different units produced during high-grade processes. Stable isotope studies of C, O, and D/H should provide insight into the sources of fluids. The style of fluid migration should also be studied. Determination of fluid to rock ratios for various units would be an initial approach.

b. The hypothesized crustal process of selective migration of certain elements from the lower to upper crust could be tested in this area if good stratigraphic and structural control would allow sampling of the same or similar units across the regional gradient.

c. Conduct chemical analyses on granulite residua and their purported partial melts as well as on individual minerals within these materials. This should provide data that can be examined for mass balance comparisons with proposed parental material, for partial melting equilibria requirements, and for distribution coefficients that are reasonable.

d. Undertake a study of regional relationships of intrusive and extrusive rocks through detailed isotopic and geochemical work. For example, how do the tonalitic gneisses in the domal gneiss terrain compare with the early and late siliceous units in the lower grade greenstone belts? These studies are crucial to understanding the effects of lower crustal processes on upper crustal processes.

e. The various metamorphic reactions, both prograde and retrograde, should be defined throughout all of the zones on the basis of changes in mineral assemblages.

f. The pressure-temperature relations for each metamorphic reaction should be determined from the mineral assemblages amenable to geobarometry and geothermometry. From this data thermal gradients should be determined for various segments of the crustal cross section.

g. Undertake Sr, Nd, and Pb isotopic studies of the differentiated pyroxenite, gabbro, diorite, syenite, and granite complex in the Kapuskasing structural zone in order to better define the petrogenesis and tectonic roles of the various igneous units.

h. Provide a petrologic and geochemical characterization of compositionally layered metasedimentary granulites to compare the potential diffusion gradients with those of quenched xenolithic granulites in volcanic rocks.

## **3. Evolution of Crust-Mantle System**

a. Complete the necessary trace element and isotopic data for a variety of rocks from the area to better understand regional mantle variations in the Archean. That is: can we define geochemical provinces

within the Archean mantle? Also, such data will provide some constraints for crustal “contaminants” that might affect younger intrusions.

b. Study the Nd-Sr-Pb-O isotopic systems of mafic dikes of various ages in order to determine the mantle source characteristics and interpret the evolution of the mantle through time as well as the interaction between the melts and the crust.

c. Further refine our knowledge of ages and isotopic data to determine how much of the Archean terrain of the Superior Province is pre-2700 m.y. old and how much is really only  $2700 \pm 50$  m.y. old.

d. Attempt to date the time of granulite metamorphism through Sm-Nd internal isochrons in garnet-bearing tonalites (i.e., date the time of garnet formation). This can be compared with zircon ages that should indicate the age(s) of the progenitors, thereby providing a more complete history of crustal evolution in the field trip area.

e. Develop a combined petrologic, geochemical, and isotopic study of the lamprophyric dikes of the Kapuskasing structure to compare with the carbonatites being studied in other parts of the shield.

f. Develop profiles of heat production and other physical parameters across the Archean continental cross section for utilization as constraints on thermal and tectonic models of Archean processes.

g. Attempt to find evidence for processes that must have acted during the pre-Archean and helped determine the nature of the future Archean crust.

## **Part B. Studies Either Underway or Proposed by Various Workers (numbers in parentheses correlate with suggestions in Part A)**

### **1. Basic Data**

George McGill intends to begin detailed structural mapping of the Wawa belt in 1984 (1a).

Gary Beakhouse indicates that the Ontario Geological Survey intends to study protoliths of supracrustal rocks within high-grade terrains by locating areas where stratigraphy can be traced across metamorphic reactions. It is possible that some of this work may be done in the general area of the field trip (1b).

Linc Hollister, who has worked extensively in Tertiary high-grade terrains, noted many similarities between the Tertiary granulite terrain in British Columbia (B.C.) and that of the Kapuskasing. There does appear to be a higher pressure of final equilibration in the Kapuskasing area as well as occurrence of anorthosite and syenogabbro that are lacking in B.C. He is now inspired to write a note to bring out these observations. He also hopes to develop a collaborative study with one of his Princeton students and John Percival (1f).

Richard Ernst is making a survey of the variations in thickness and trends of dikes of various ages across the central and southern Superior Province to develop maps of changes in crustal depth-of-exposure in Archean and younger times (1g).

Tom Feininger may be charged with refining the gravity interpretation of the Kapuskasing structural zone by the Earth Physics Branch in Ottawa (part of suggestion 1h).

Dave Fountain, in collaboration with Matt Salisbury, hopes to measure sonic velocities at high P and T for samples across the field trip area. He is also interested in collaborative efforts with others to measure thermal conductivity, Curie temperatures, magnetic susceptibility, and electrical conductivity on the same suite of samples (1i).

Lew Ashwal, in collaboration with Paul Morgan, plans to measure heat production from a suite of samples collected across the field trip traverse and develop a heat production profile across the Kapuskasing-Wawa crustal section. This is crucial to thermal modeling of continental areas (1c and 3f).

John Percival has immediate interest in correlating petrologic properties of rocks with their geophysical parameters (1i).

## 2. Lower Crustal Processes

John Valley plans to analyze oxygen isotopes from the Shawmere anorthosite to determine possible fluid interactions during granulite facies metamorphism (2a).

Darrell Henry, Roberta Rudnick, and Lew Ashwal are completing a study of fluid inclusions in the Kapuskasing rocks. This involves not only the changes in P and T with time, but also will include general calculations on the role of carbonic fluids in granulite grade metamorphism (2a).

Denis Shaw and Marilyn Truscott are presently studying Boron distribution in Precambrian cherts, Fe formations, and shales in order to determine the B content of the Precambrian oceans (they will also study Cd, Gd, and Sm in the near future). This is to be expanded to determine the behavior of B during deep crustal metamorphism through analyses of granulite facies whole rocks. The objective is to determine whether original B in igneous and sedimentary rocks is immobilized by formation of an erosion resistant mineral like tourmaline and lost from the sedimentary cycle or is remobilized in a fluid phase (partly 2a and partly 2b).

Ross Taylor and Roberta Rudnick plan to compare the chemistry of greenschist facies basalts and tonalites with mafic and tonalitic gneisses of the granulite facies, as well as trace element and isotope geochemistry of partial melts and restites within granulite facies rocks (2a, 2b, and 2c).

Gil Hanson, in collaboration with John Percival, plans to undertake reconnaissance studies of the conditions of partial melting in granulites. Gil will probably put a graduate student to work on the project. Depending on the results, this could become a much larger project in the future.

Larry Haskin intends to commence reconnaissance analyses on samples of granulite and their potential partial melts, as well as on minerals within these materials. The results will be examined for overall mass balance to compare with related materials in the literature, for consistency with partial melting equilibria, and for distribution coefficients. Depending on the results of this reconnaissance, he might propose a detailed study of one or two single outcrops from which many samples would be extracted to establish chemical trends and extent of heterogeneity within individual units (2b and 2c).

Ross Taylor and Roberta Rudnick also plan to use trace element chemistry to study the source of high-Al basaltic compositions in the gneisses. Do they represent mantle derived melts or restites? In a related study they plan to study anorthosite origin, i.e., are they partial melts of the mantle or of lower crustal high-Al basalt compositions (2c)?

Klaus Schulz and Paul Sylvester, in collaboration with Kodjo Attah, are working on the geochemistry and petrogenesis of the Wawa area volcanic rocks in order to determine the sources of the melts and the relationships between them (a necessary part of 2d and 3a).

Paul Sylvester and Bill Phinney, in collaboration with Paul Studemeister, will undertake petrologic and geochemical studies of felsic rocks within the Michipicoten greenstone belt in an attempt to determine the sources and crystallization history of granitic and trondhjemitic magmas in greenstone belts. This should lead to a better knowledge of the overall mass balance of greenstone belt source materials (part of suggestion 2d as well as 1i).

George Tilton plans to study Pb isotopes from the Shawmere anorthosite in order to develop a Pb/Pb isochron for dating purposes and compare initial Pb with that of other areas in the nearby Shield (3a and 3c). He will also study the isotopic systematics of granulites and tonalites from the Chapleau-Foley region to gain some idea of the nature of possible crustal contamination that might affect the younger rocks that he has studied in the area. The Pb work will come first; Sr and Nd data may follow later (3a).

Richard Ernst is conducting a geochemical study of a 1.1-b.y.-old mafic dike that extends for >500 km from the Kapuskasing structural zone to the Grenville Front. This should aid in models of mantle heterogeneity, crustal contamination, and tectonics (part of suggestion 3a).

Elaine Padovani, Ken Collerson, and Ken Card will undertake a petrological, geochemical, Sr-Nd-Pb isotopic study of lamprophyre dikes in the Kapuskasing structural zone to compare the isotopic systematics with those of carbonatites being studied elsewhere (3a and 3e).

Rick Carlson and Steve Shirey plan to choose among topics 3b and 3d or a combination of 3a and 2d as applied to the tonalitic gneisses from the Wawa domal gneiss terrain and the intrusive siliceous units in greenstone belts.

John Percival has immediate interests in carrying out studies of zircon geochronology in metamorphic rocks (part of 3c and 3d).

Marilyn Truscott hopes to study the lamprophyre dikes and their (lower crustal?) inclusions in order to compare them with their younger counterparts for changes, if any, in source and manner of emplacement (3e).

## Part C. Summary of Discussion

### Publications

Several types of publications were considered to be desirable and include the following:

1. Abstract volumes and appropriate notes to be published as LPI Technical Reports (as soon as possible after meetings).
2. Special issues or sections of issues on specific themes or topics, particularly those resulting from workshops, topical conferences, or special sessions at large meetings. *Journal of Geophysical Research* seemed to be a favorite journal for publication of groups of papers.
3. Compilations of reprints from above categories of publications should be available for panels, advisory groups, program managers, and appropriate persons at NASA headquarters or other organizations that need justification for support of this program.
4. Some type of summary volume at the end of the program (10 years?) is favored by several persons.
5. Provide writeups of workshops, conferences, etc., to general circulation journals (*EOS*, *Geology*, *Geotimes*, *Geoscience Canada*, etc.) as soon as possible after the meetings.

### Conferences, Workshops, and Meetings

1. Some sort of regularity should be established for the ECG program. For example, the advisory group should establish a schedule to include all of the elements of the program and schedule, at least tentatively, for the next two or three years the workshops, conferences, special sessions at meetings (GSA, GAC, AGU, etc.), deadlines, and possible publications. Several persons requested that copies of such a schedule be distributed to interested parties as soon as possible to allow for coordination and avoid conflicts, overlap, or repetition. For example, in 1984 the IGC meeting in Moscow has planned a field trip to Aldan Shield and in 1986 the GAC in Ottawa is planning a deep crust session convened by Dr. T. M. Gordon, Geological Survey of Canada.
2. A series of topics could be coordinated between workshops, conferences, special sessions, and publications. These should be process-oriented or thematic, not regional. For example, a field trip for "lower crustal processes" followed a year or two later by a special session on the same topic could form the basis of a special issue of a journal.
3. Try to avoid meetings that limit the number of participants. Aim for conferences, workshops, special sessions, etc., that are open to all potential ECG contributors.
4. Field workshops were considered to be an excellent means of exchanging ideas, developing new approaches, and coordinating research plans. It was nearly unanimously accepted that another such workshop be organized for the summer of 1985. Werner Weber of the Manitoba Geological Survey suggested that he might look into the requirements and logistics for a similar workshop organized

around the topic of “lower crustal processes” in the Pikwitonei province of northern Manitoba. There was extensive enthusiasm for such a trip.

### **Miscellaneous Items**

1. There should be an advisory group that would include representatives from other groups with mutual interests, such as USGS, Archean Geochemistry, Canadian organizations, Lithosphere Committee, etc. This should help avoid conflicts and promote communications.
2. For additional sources of funding for meetings, student support, field work, etc., there may be several possibilities, including the GSA Penrose fund for conferences, the GAC Foundation in Canada, the Ontario Geological Survey, which supports certain types of research and field studies, and others. It might be worthwhile to compile a list of such sources.
3. A newsletter should be sent to interested persons with some degree of regularity.
4. Because the Early Crustal Genesis Program will have a rather limited capability for financial support of proposals, a major effort should be made to coordinate conferences, workshops, special sessions at meetings, and publications in conjunction with other groups having related interests.

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# ABSTRACTS



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THERMAL CONSTRAINTS ON HIGH-PRESSURE GRANULITE METAMORPHISM OF SUPRACRUSTAL ROCKS L.D. Ashwal, P. Morgan, and W.W. Leslie\*, Lunar and Planetary Institute, 3303 NASA Road 1, Houston, TX 77058. \*also at Dept. Earth & Space Sci., UCLA, Los Angeles, CA 90024.

We examine here the circumstances leading to the formation and exposure at the Earth's surface of supracrustal granulites. These are defined as sediments, volcanics, and other rock units which originally formed at the surface of the Earth, were metamorphosed to high-pressure granulite facies ( $T = 700-900^{\circ}\text{C}$ ,  $P = 5-10$  kbar), and reexposed at the Earth's surface, in many cases underlain by "normal" thicknesses of continental crust (30-40 km). Examples are numerous, and are represented by Archean through Tertiary occurrences (Table 1). Three stages in the formation of such rocks must be accounted for: transport from the surface to depths of 15-30 km, heating to  $700^{\circ}\text{C}$  or more, and reexposure at the surface.

Tectonic underthrusting is the most plausible mechanism to transport surface rocks to 15-30 km depths. This can be achieved either by continental underthrusting and consequent double-thickening to 60-80 km (4, 26), or by underthrusting of thin plates, with only minor increase in crustal thickness (18). Each of these cases places distinctly different constraints on possible thermal histories for the supracrustal rocks, as discussed below. Although burial of supracrustal rocks to 15-30 km by continuous sedimentary-volcanic loading is a possibility, it must be a remote one, based on isostatic considerations.

There are several mechanisms by which such rocks can be re-exposed at the surface. Uplift is a natural consequence of isostatic readjustment in thickened continental crust, and where double-thickening has occurred by underthrusting, the top of the underthrust plate can be reexposed by subsequent erosion. The granulite exposures in the Massif Central have been suggested to have formed in this way (1). Large-scale isostatic movements, however, cannot account for uplift to the surface of thin slivers underthrust below continental crust of normal (35-40 km) thickness. In this case, tectonic mechanisms such as deep reverse faults would be necessary to account for surface exposures of supracrustal granulites. This usually results in a crustal cross-section, with a continuous increase in metamorphic grade from greenschist to granulite toward the fault (8). Examples of granulite terranes thought to have developed in this way include the Ivrea Zone of the Alps (16), and the Kapuskasing Structural Zone of Ontario (19). Thus, although particular tectonic conditions may be required for the burial and subsequent reexposure of supracrustal rocks of granulite grade, these conditions are easily explained in the framework of plate tectonics.

The heating step is perhaps the most difficult to account for. Intrusion of magmas at temperatures greater than  $1000^{\circ}\text{C}$  could provide the necessary heat for granulite metamorphism, but most granulite terranes do not contain the requisite volumes of mafic intrusives to explain the heating directly by mantle-derived magmas (25). Hot crustal melts such as tonalites could provide a solution for terranes containing such materials, such as for the Tertiary granulites of the Coast Ranges of British Columbia (10, 11). The tonalitic melts may be produced by anatexis or hybridization (29), but the ultimate heat

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Table 1. Compilation of Some Supracrustal Granulite Occurrences

LOCATION	ROCK UNITS	P-T CONDITIONS*	AGE (OF META.)	REF.
<u>ARCHEAN</u>				
Beartooth Mtns., Montana	Basalts, Pelites, Ironstones	7 kbar 650-750°C	3400 Ma	9
Enderby Land, Antarctica	Pelites, Ironstones, Calc-Silicates, Marbles	8-10 kbar 900-950°C	3000 Ma	24
Malene, West Greenland	Marbles, Calc-Silicates Metavolcanics	7-9 kbar 650-850°C	3000 Ma	6
Lewisian, Scotland	Pelites, Aluminous & Siliceous Sediments	10-13 kbar 800-860°C	2700- 2900 Ma	15,28
Sierra Leone	Pelites, Iron Formations	6-9 kbar 720-820°C	2800 Ma	21
Pikwitonei Domain, Manitoba	Pelites, Iron Formations	10-11 kbar 900-1000°C	2400- 2800 Ma	27
Kapuskasing Struc- tural Zone, Ont.	Metabasalts, Marly Sediments	6-8 kbar 700-800°C	2600- 2700 Ma	19
<u>PROTEROZOIC</u>				
Broken Hill, Australia	Pelites, Calc- Silicates	5-6 kbar 650-800°C	1700 Ma	20
Adirondacks, New York	Marbles, Quartzites Semi-Pelites	6-8 kbar 650-750°C	1000- 1100 Ma	5
<u>PHANEROZOIC</u>				
Ivrea Zone, No. Italy	Pelites, Marbles, Migmatites	8-11 kbar 700-820°C	450 Ma	12,22
Massif Central, France	Pelites, Meta- basalts	11 kbar 800°C	350- 400 Ma	1,2
Coast Ranges, British Columbia	Metabasalts, Calcic & Aluminous Sediments	5-8 kbar 750-850°C	62 Ma	10,11

\* In some cases P-T estimates were determined from rock units adjacent to or in association with the supracrustal granulites.

source for such melts is probably mantle-derived magmatism.

Other heating mechanisms are somewhat constrained by the way in which the supracrustals are reexposed after metamorphism. Consider first exposure by erosion and isostatic readjustment of crust double-thickened by underthrusting. Several models of this type have been published which attempt to predict the thermal histories of various points within the crust during erosion following an "instantaneous" underthrusting event (3, 7, 25). Most of these models assume that the supracrustal rocks near the thrust are heated primarily during the thermal relaxation of the perturbed temperature profile (commonly a "sawtooth" profile) following underthrusting. The effects of erosion are to increase the temperature at any particular depth, but to decrease the maximum temperature acquired by any particular rock unit during thermal relaxation as it rises to the surface. Maximum possible temperatures will be attained by any particular rock unit if thermal relaxation is complete before erosion occurs. Although we recognize that this condition is geologically unrealistic, we use it to demonstrate that granulite metamorphism of supracrustal rocks in the middle of double-thickened crust by conductive heating alone is highly improbable.

A range of possible steady-state geotherms in double-thickened crust are shown in Fig. 1. Most of the possible geotherms either fail to attain sufficient temperatures at mid-crustal levels to cause granulite metamorphism of the underthrust supracrustal rocks and/or exceed the probable solidus temperature near the base of the crust, suggesting that melts would form, and that the assumption of a conductive geotherm and conductive thermal relaxation is invalid. Only geotherms generated assuming extreme upper crustal enrichment of heat production with little or no mantle heat flow (e.g. E and F, Fig. 1) can produce mid-crustal granulites without lower crustal melting, and these conditions must be regarded as highly improbable. Thus, we conclude that conductive heating alone cannot produce granulites from supracrustal rocks in the middle of a double-thickened crust. This mechanism could, however, be supplemented by magmatic heating as discussed above, and could apply to granulite terranes containing substantial volumes of crustally derived granitic melts produced during the metamorphism (21).

A possible mechanism to account for granulite terranes which lack evidence for magmatic activity during metamorphism involves underthrusting of thin ( $< 5-10$  km) slivers of supracrustals. Conductive heating alone could produce granulite metamorphism of these supracrustal rocks without necessarily melting the lowermost crust. These rocks must then be exposed by a subsequent tectonic overthrusting event, as they cannot be brought to the surface isostatically during erosion.

Conductive heating in a double-thickened crust to produce granulite metamorphism at mid-crustal levels can be augmented or dominated by two mechanisms: (i) preheating of the crust prior to underthrusting, and (ii) shear heating along the thrust. Preheating of the overthrust crust is likely to occur by arc magmatism if significant subduction preceeds overthrusting. Preheating may also occur associated with pre-orogenic granites (4), which may be associated with delamination of the mantle portion of the lithosphere prior to overthrusting. Thus if the lower portions of the overthrust plate are sufficiently hot ( $800-1000^{\circ}\text{C}$ ), thermal relaxation could produce

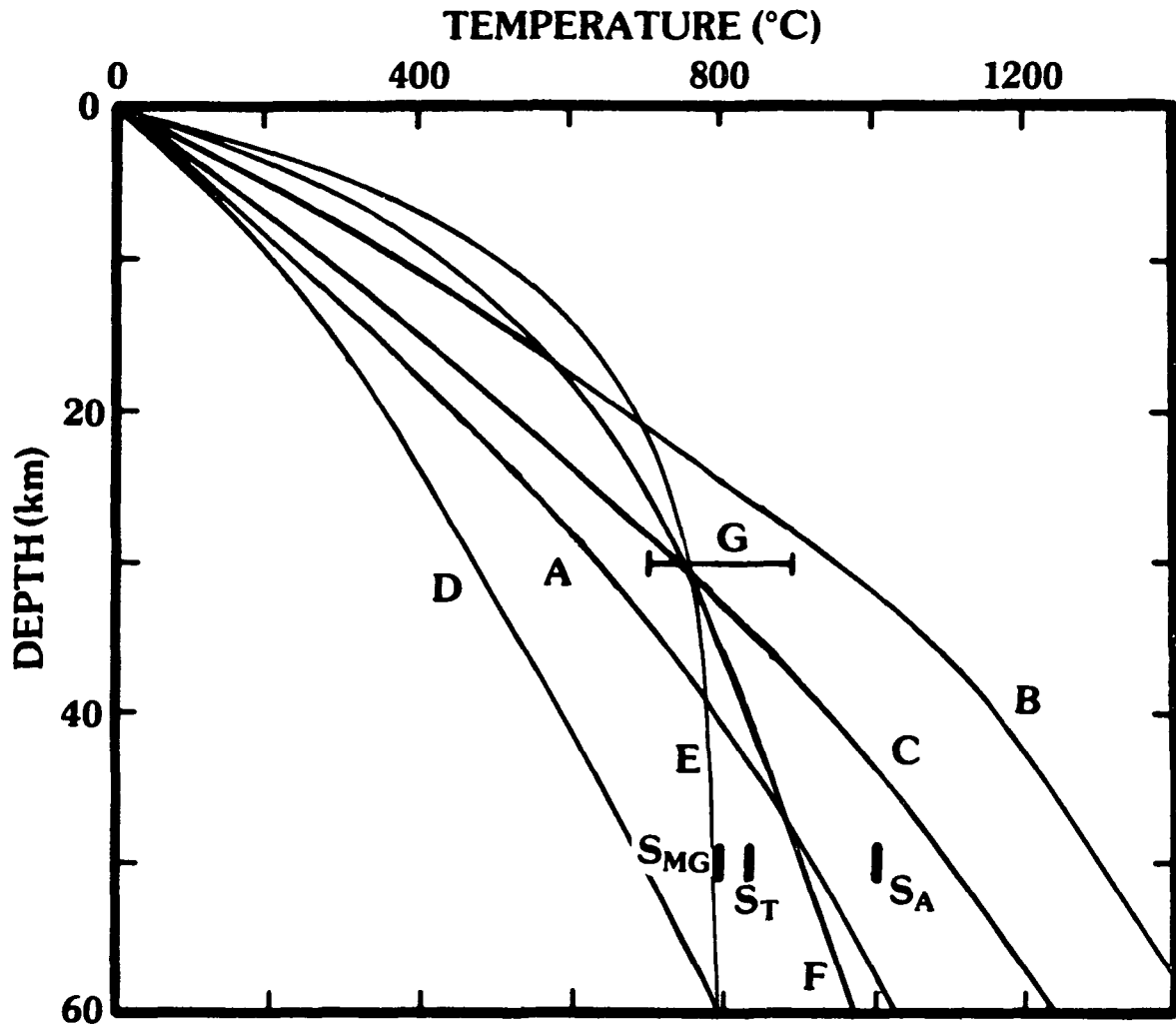


Fig. 1. Possible equilibrium geotherms in double-thickened crust composed of two 30 km thick plates to illustrate the relationship between mid-crustal and Moho temperatures. Heat producing elements, where present, are assumed to be exponentially distributed with depth through the plates, with a depth distribution parameter of 10 km (e.g. 13). Curves A, B, and C were calculated assuming identical heat production in both upper and lower plates. In curve A, a heat production,  $A_s$ , of  $2.0 \mu\text{W m}^{-3}$  was assumed at the top of each plate with a mantle heat flow,  $Q_m$ , of  $27 \text{ mW m}^{-2}$  (typical shield value); for curve B,  $A_s = 4.0 \mu\text{W m}^{-3}$ ,  $Q_m = 27 \text{ mW m}^{-2}$ ; and for curve C,  $A_s = 2.0 \mu\text{W m}^{-3}$ ,  $Q_m = 36 \text{ mW m}^{-2}$ . Curves C, D, and E were calculated assuming all heat production to be concentrated in the upper plate. For curve D,  $A_s = 4.0 \mu\text{W m}^{-3}$ ,  $Q_m = 27 \text{ mW m}^{-2}$ ; for curve E,  $A_s = 20 \mu\text{W m}^{-3}$ ,  $Q_m = 0$ ; for curve F,  $A_s = 14.4 \mu\text{W m}^{-3}$ ,  $Q_m = 17 \text{ mW m}^{-2}$ . A uniform crustal thermal conductivity of  $2.5 \text{ W m}^{-1} \text{ K}^{-1}$  was assumed. The granulite temperature field at mid-crustal levels (30 km) is marked by G.  $S_{MG}$ ,  $S_T$ , and  $S_A$  are solidus temperatures at 50 km depth for muscovite granite, tonalite, and amphibolite, respectively, assuming  $\text{H}_2\text{O}$  present sufficient for formation of muscovite, biotite, and hornblende, but without excess  $\text{H}_2\text{O}$  (from ref. 29). If thinner plates are considered, the geotherms can be scaled linearly to yield similar mid-crustal/Moho temperature relationships. It is not proposed that very high temperatures are likely in the lower plate - the curves are used to illustrate the point that in a purely conductive regime, granulite metamorphism temperatures are highly unlikely to be generated at mid-crustal levels without significant melting in the lower plate.

supracrustal granulites in the top of the underthrust plate without initiating melting below. Alternatively, or in addition, shear heating on the thrust may also occur. It is usually assumed that this shear heating is of minor importance (e.g. 5), but if high frictional stresses can be maintained on the thrust, shear heating can produce melting (17). Thus, temperatures sufficient for granulite metamorphism may be generated in the region of the thrust plane. The efficiency of shear heating is still a subject of much uncertainty (e.g. 14, 23), but should this mechanism produce sufficient temperatures for granulite metamorphism, the metamorphic isograds would be strongly controlled by faults.

In summary, we propose 5 possible heating mechanisms to account for granulite metamorphism of supracrustal rocks:

1. Magmatic heating, (a) from mantle-derived melts, or (b) from anatectic products of mantle-derived melts.

2. Thermal relaxation of perturbed temperature profiles following underthrusting and double-thickening of continental crust. This necessarily results in crustal melting of the lower plate unless crustal heat generation and mantle heat flow are very low. Granulite terranes formed in this way may be indistinguishable from those produced by mechanism 1(b).

3. Thermal relaxation after underthrusting of thin slivers of supracrustal rocks below continental crust of "normal" thickness (30-40 km). This avoids anatexis during metamorphism, but requires a subsequent tectonic event to elevate the granulites to the surface.

4. Major preheating of the upper plate (for example by arc magmatism or pre-orogenic granites) prior to underthrusting.

5. Shear heating caused by high frictional stresses along the thrust plane.

It is clear that granulite metamorphism of supracrustal rocks is not the simple consequence of continental collision. Magmatism and/or subsequent tectonic events almost certainly play a fundamental role in the heating and reexposure of the metamorphosed supracrustal rocks. Identification of the thermal histories of such terranes requires detailed input from field studies and geochronology. In particular, the relationship of spatially associated granitic rocks to the metamorphic event(s) must be carefully evaluated.

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# IS THE KAPUSKASING STRUCTURE THE SITE OF A CRYPTIC SUTURE?

Kevin Burke, Lunar and Planetary Institute, 3303 NASA Road 1, Houston, TX, 77058

In 1968 J. T. Wilson suggested that the Circum-Ungava suture zone continued through the Kapuskasing to join the Penokean fold belt (Fig. 1) implying that the Kapuskasing marked the site of what has since come to be known as a "cryptic suture" (e.g., Burke and Dewey, 1973). Later workers, however, including both those who followed Wilson in relating the Circum-ungava structure to ocean opening and closing (e.g., Gibb and Walcott, 1971; Burke and Dewey, 1973) and those who recognized only rift features within it (e.g., Baragar and Scoates, 1981) have preferred to extend the Circum-Ungava structure through the Thompson (Nelson) zone of Manitoba (Fig. 1).

Now Percival and Card (1983, see especially Fig. 2) have demonstrated that the Kapuskasing structure involves substantial thrusting of deep continental crustal rocks over shallower continental rocks. Because this is a process typically (though not uniquely) associated with continental collision, there may be a case for looking again at Wilson's original suggestion.

Problems arise in attempting to reconcile Wilson's idea with data from more recent studies:

## 1) Could the Kapuskasing and the Thompson belt both mark sutures of about 1700 Ma age?

Geometric relations similar to those shown in Fig. 1A with the subsurface extension of the Thompson belt meeting the Circum-Ungava and Kapuskasing structures in a T-junction are known from numerous places around the world where collisional mountain belts join (e.g., in the Damarides of Namibia) these junctions have been called "Aral knots" (Sawkins and Burke, 1980, Fig. 1) from the type example where the Urals meet the Hercynian fold-belts of Europe and Kazakhstan.

## 2) Why is there no age difference across the Kapuskasing if it does mark the site of a continental collision?

For the Kapuskasing to mark a collision without age resetting (except for that contemporary with information of the Ivanhoe Lake cataclastic zone, Percival and Card, 1982) would require that the western Superior province 'docked' gently against the east without producing any of the effects usually recognized at collision.

## 3) Why is there no offset of the Superior subprovinces across the Kapuskasing?

Percival and Card (1983, p. 326) report that there is no major offset of the Abitibi - Opatika boundary across the Kapuskasing. It would be remarkable for the belts to have been sutured together in a matching configuration.

If Wilson's (1968) interpretation of the Kapuskasing structure is valid, the suturing involved would have had to be remarkably cryptic, but because we know very little of the properties of suture zones (except for their complexity, Dewey, 1976) and because Wilson has proved right so often in the past, it would seem that his hypothesis merits further testing.

Percival and Card's perception (1983) that the timing of tectonic events in the Kapuskasing matches that of events elsewhere in the Canadian shield opens the way for other possible explanations of the origin of the Kapuskasing structure. The age of the Ivanhoe Lake cataclastic zone in the Kapuskasing (reported as 1720 Ma, Percival and Card, 1983) is similar to that of an event which has been interpreted as collisional on the Thompson front and this leads to the suggestion that the Kapuskasing may be an isolated upthrust area within one of two colliding continents, possibly comparable to the present-day Tien Shan (Fig. 2B), a 5 km high upthrust range produced in the active Indian-Asian continental collision (Molnar and Tapponnier, 1975).

The Tien Shan uplift is: 1) contemporary with collision and suturing 1000 km away; and 2) isolated within an area that has not been uplifted; and 3) apparently not associated with igneous activity (i.e., it is unlikely to have reactivation and resetting of isotopic systems at depth).

A significant difference between the active Tien Shan and the ancient Kapuskasing structure is that the former occurs on the side of the Indus suture zone which carried an Andean arc before continental collision (the "hot" side) while the Kapuskasing occurs on the opposite side (the "cold" side). If this proves to be a significant difference, then the isolated uplifted areas of Peninsular India (e.g., the Nilgiri Hills, about 2 km high), may prove a better analogue of the Kapuskasing structure. A representation of how the Indian-Asian and Superior-Churchill collisions might be analogous is sketched in figure (2).

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# THE KAPUSKASING STRUCTURE

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K. Burke

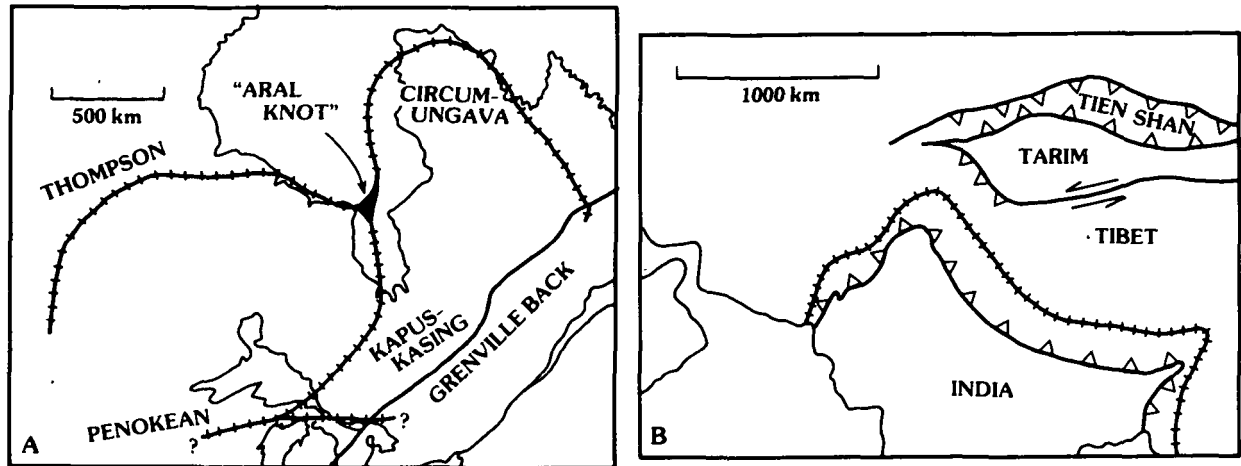


Fig. 1 (A). Possible distribution of suture zones in the Canadian shield. Wilson's (1968) suggestion that the Kapuskasing marks a cryptic extension of the Circum-Ungava suture and Gibb and Walcott's (1971) suggestion that the extension is in the Thompson belt are both assumed valid. The kind of suture T-junction illustrated is common and is known (from the type example) as an Aral knot.

(B) Sketch illustrating the disposition of the Tien Shan (an isolated 5 km high upthrust mountain range in Asia) with respect to the Himalaya and the Indus-Yarling Dzungpo suture zone. A continental collision on the Thompson front 1700 Ma ago might have generated the Kapuskasing upthrust zone as a structure similar to the Tien Shan of today. This explanation seems more compatible with the evidence than the idea that the Kapuskasing marks a cryptic suture-zone.

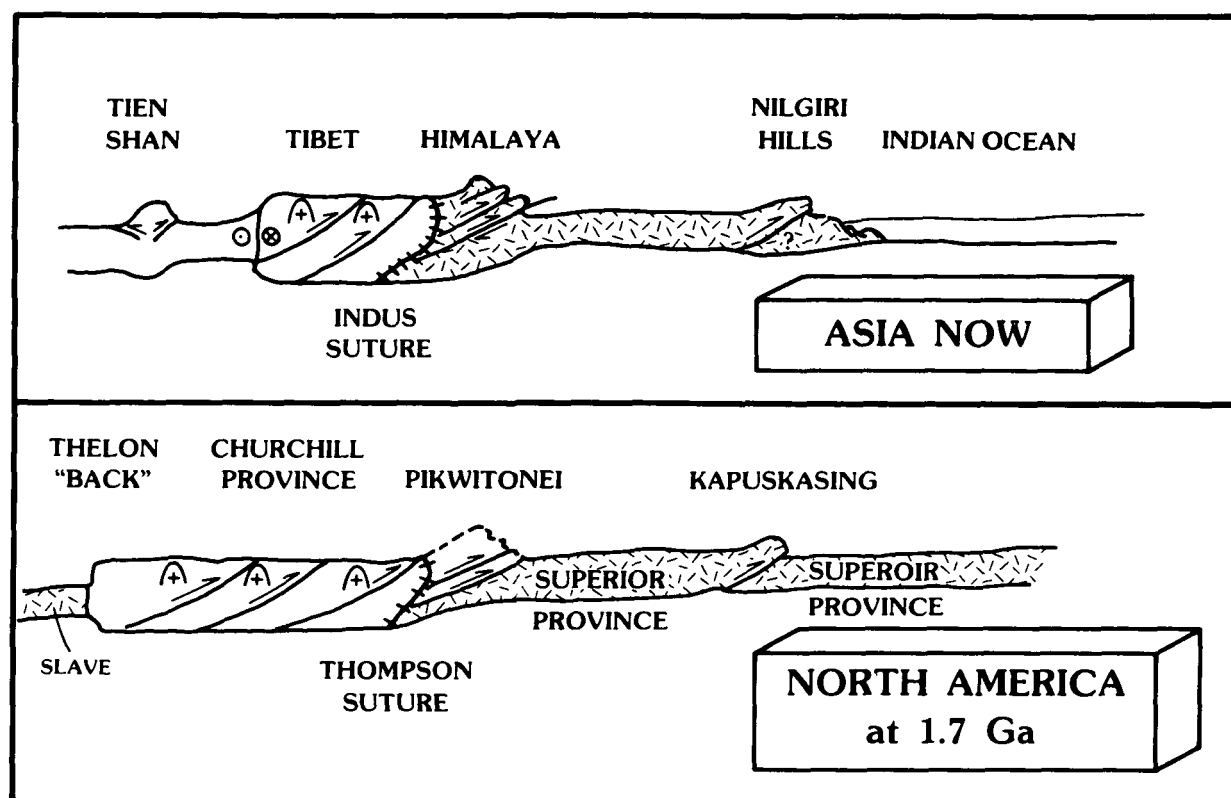


Fig. 2. A comparison of the structure of Asia now (illustrated in the upper sketched cross-section) and North America 1.7 Ga ago (lower sketched cross-section) to illustrate that just as uplift of an isolated area in Peninsular India (e.g., the Nilgiri Hills) may be related to the Himalayan collision so uplift forming the Kapuskasing structure may have been related to a comparable collision along the Thompson front. On this interpretation Peninsular India and the Superior Province play similar roles at collision; the Pikwitaneí subprovince marks the site of an eroded analogue of the Himalayan range. The Indus and Thompson suture zones are comparable, the Churchill province is an analogue of Tibet with thickened continental crust and reactivation. The Thelon 'back' marks the boundary between thickened and unthickened continental crust and is analogous to the northern boundary of the Tibetan plateau. The Tien Shan, an alternative analogue of the Kapuskasing structure, is shown on the Asian cross-section.

Sm-Nd ISOTOPIC SYSTEMATICS OF THE ANCIENT GNEISS COMPLEX, SOUTHERN AFRICA. R. W. Carlson<sup>1</sup>, D. R. Hunter<sup>2</sup>, and F. Barker<sup>3</sup>; <sup>1</sup>Department of Terrestrial Magnetism, 5241 Broad Branch Rd., Washington, D.C., 20015, U.S.A.; <sup>2</sup>University of Natal, Pietermaritzburg, 3200, Natal, South Africa; <sup>3</sup>U. S. Geological Survey, 4200 University Dr., Anchorage, Al., 99508

The igneous core of southwestern Africa's Kaapvaal craton consists of the Onverwacht Group of mafic to ultramafic volcanics of the Barberton greenstone belt and a complex gray gneiss terrain called the Ancient Gneiss Complex (AGC) (e.g. 1,2). Until recently, precise geochronologic information for these two units has been difficult to obtain due to the effects of post-formation metamorphism. Even the assignment of relative ages between the AGC and the Onverwacht is complicated by the lack of direct contact between these two units in the field(1).

Recently, by applying the Sm-Nd radiometric system, Hamilton et. al.(3) determined a whole-rock age of  $3,530 \pm 50$  Ma for the lower ultramafic unit of the Onverwacht Group. Compared to this age, Rb-Sr dates for gneisses of the oldest unit of the AGC, the Bimodal Suite (BMS), tend to be slightly younger ( $3,200$ – $3,300$  Ma; 4). However, these Rb-Sr ages most likely reflect later metamorphic episodes rather than the emplacement ages of the inter-layered metabasalts and tonalite-trondhjemite gneisses that make up the BMS. Based on a correlation between initial  $^{87}\text{Sr}/^{86}\text{Sr}$  and age of individual gneiss units within the AGC, Davies and Allsopp (4) suggested an emplacement age of about  $3,400$  Ma for the AGC parental materials, some  $100$  Ma younger than the Onverwacht volcanics. In contrast, Barton et al.(5) reported a Rb-Sr whole rock age of  $3,555 \pm 111$  Ma for the BMS placing its formation at about the same time as the Onverwacht Group.

In order to shed some new light on the question of the absolute and relative ages of the AGC and Onverwacht Group, a Sm-Nd whole-rock and mineral isochron study of the AGC was begun. At this point, the whole-rock study of samples from the BMS selected from those studied for their geochemical characteristics by Hunter et al.(6) has been completed. We discuss here these results and their implications for the chronologic evolution of the Kaapvaal craton and the sources of these ancient rocks.

Sm-Nd data for samples of the BMS (Table 1) precisely define a line on the isochron diagram shown in the figure. The line corresponds to an age of  $3,417 \pm 34$  Ma with an initial  $^{143}\text{Nd}/^{144}\text{Nd} = 0.508149 \pm 31$  or initial  $\epsilon_{\text{Nd}} = +1.1 \pm 0.6$  (using the "bulk-earth" Sm-Nd parameters of Jacobsen and Wasserburg(7)). All data points lie within analytical uncertainty of the best fit line with the exception of the data for sample SWZ-6 which lies below the isochron by only 5 parts in  $10^5$ . The excellent colinearity of these data is surprising given the wide variation in chemical composition of the samples from siliceous gneiss (SWZ-5;  $\text{SiO}_2 = 76\%$ ) to tonalite (SWZ-6;  $\text{SiO}_2 = 66\%$ ), diorite (SWZ-3;  $\text{SiO}_2 = 57\%$ ), and metabasalt (SWZ-10 and 12;  $\text{SiO}_2 = 49\%$ ). Even the data for the stratigraphically younger quartz-monzonite gneiss, SWZ-4, lie on the same line defined by the remainder of the data. If the data for SWZ-4 are left out of the line regression, the isochron

shifts to only a slightly older age of  $3.45 \pm 0.04$  Ga with the initial  $\epsilon_{Nd}$  increasing by only 0.1 unit.

Because of the excellent colinearity of the Sm-Nd data, the age indicated is interpreted as the time of extraction of the parental materials of the BMS from a common, isotopically homogeneous, source. The positive initial  $\epsilon_{Nd}$  for the AGC indicates that its source was in the mantle and not in a much older, LREE enriched, crustal section. This is not to say that all the varied chemical species that make up the AGC were derived directly from the mantle in a single igneous event. Rather, the Sm-Nd data allow for a variety of petrogenetic mechanisms, from direct partial melting of the mantle for the basaltic components to anatectic melting of slightly older crustal materials for the more silicic rocks as long as these events occurred within the time interval specified by the isochron.

The age determined for the BMS is distinct and some 100 Ma younger than that of the Onverwacht volcanics(3) in accord with the suggestion of Davies and Allsopp (4). Though the similar initial isotopic compositions of the AGC and Onverwacht allow for some of the more silicic members of the AGC to have been derived by remelting of a basaltic crust of Onverwacht age, the presence of basaltic members in the BMS also indicates a significant contribution to the AGC from the mantle. Thus the Kaapvaal craton appears to have originated by at least two episodes of mantle derived mafic volcanism occurring over a period of about 100 Ma.

The positive initial  $\epsilon_{Nd}$  of the BMS, and its agreement with that determined for the Onverwacht Group(3), shows the presence of a relatively homogeneous mantle source for the oldest units of the Kaapvaal craton. This source appears to have been relatively depleted in the LREE for a considerable time prior to the formation of the AGC and the Onverwacht volcanics. The increasing occurrence of positive initial  $\epsilon_{Nd}$  values for the oldest crustal rocks (e.g. 8-11) implies either that differentiation events within the earth occurred well before (i.e. several hundred million years before) the time of preservation of the oldest observed crustal sections, or that a "chondritic" model for the evolution of the Sm-Nd system of the bulk-earth is not appropriate. The answer to this question carries with it important consequences for models of the bulk composition and early evolution of the earth. However, because of the possible homogenizing effects of convective mixing within the mantle over earth history, further constraints on the early geochemical evolution of the earth clearly must be sought in more complete and precise isotopic data for the ancient rocks of the continental crust.



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Table 1: Sm-Nd Results for samples of the Bimodal Suite

Sample	Sm <sup>a</sup>	Nd <sup>a</sup>	<sup>147</sup> Sm/ <sup>144</sup> Nd <sup>b</sup>	<sup>143</sup> Nd/ <sup>144</sup> Nd <sup>c</sup>
SWZ-4	2.39	16.1	0.08982	0.510293±18
SWZ-6	3.20	20.0	0.09661	0.510403±25
SWZ-5	9.66	45.5	0.1282	0.511146±20
SWZ-3	2.07	7.89	0.1584	0.511822±24
SWZ-10	1.68	5.21	0.1953	0.512668±29
SWZ-12	0.772	2.25	0.2078	0.512946±29

- a) Concentrations expressed in ppm. Sm and Nd blanks of 20 pg. and 80 pg. respectively, are negligible for the sample sizes analysed.
- b) Determined with tracers calibrated with AMES Sm and Nd metal standard solutions and cross-checked against the CIT n(Sm/Nd)  $\beta$  standard. Uncertainty  $\sim 0.1\%$ .
- c) Measured as NdO<sup>+</sup>, fractionation corrected to  $^{146}\text{NdO}/^{144}\text{NdO} = 0.722251$  ( $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$ ). Data reported relative to a value of  $^{143}\text{Nd}/^{144}\text{Nd} = 0.511860$  for the La Jolla Nd standard.

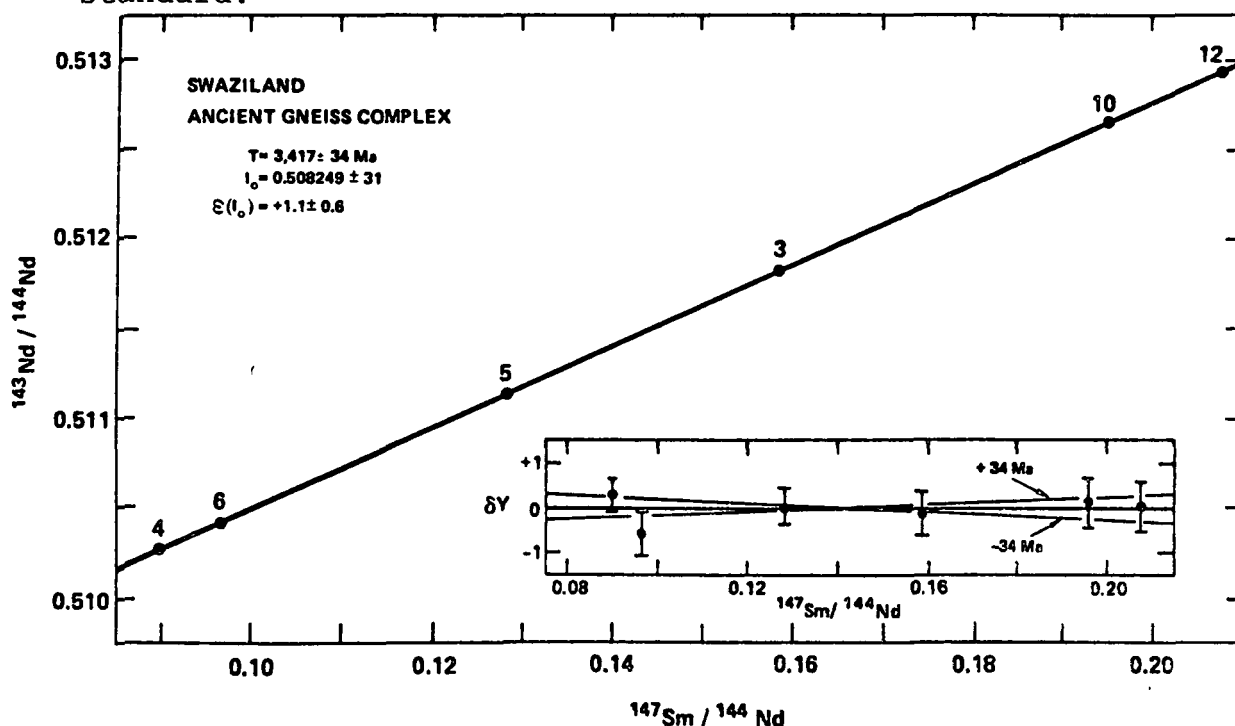


Figure 1: Sm-Nd isochron diagram for samples of the Bimodal Suite. Inset shows deviations ( $\delta Y$ ) in parts in 10,000 of the data from the best fit line.

ION MICROPROBE ZIRCON GEOCHRONOLOGY OF THE UIVAK GNEISSES: IMPLICATIONS FOR THE EVOLUTION OF EARLY TERRESTRIAL CRUST IN THE NORTH ATLANTIC CRATON. Kenneth D. Collerson, Research School of Earth Sciences, Australian National University, P.O. Box 4, Canberra, A.C.T. 2600, Australia.

*Introduction.* Geochronological studies of high-grade metamorphic rocks from Labrador and Greenland using conventional Rb-Sr, Pb-Pb, U-Pb, Sm-Nd and Lu-Hf isotopic techniques [1-13] have provided important information concerning: (1) the distribution of early terrestrial crust in the North Atlantic Craton (NAC), (2) the isotopic character of the mantle from which this crust was derived, and (3) the response of early crustal areas to subsequent metamorphic events. Nevertheless, interpretations of such isotopic data, in particular the discrimination between protolith and metamorphic ages, as well as the significance (in terms of crustal residence times) of initial  $^{87}\text{Sr}/^{86}\text{Sr}$  (ISr) values are commonly equivocal. The Sm-Nd and Lu-Hf isotopic systems have provided a potentially useful additional constraint for rationalizing these interpretations. This is because the REE's are generally considered to be less prone to metamorphic disturbance than Rb and Sr. Unfortunately, "cogenetic" suites of high-grade gneisses commonly exhibit little variation in degree of REE fractionation, hence they exhibit limited ranges in Sm/Nd and Lu/Hf. As a result, isochrons are generally of poor quality and both petrogenetic and geochronological conclusions are commonly strongly model dependent.

Although greater accuracy can now be achieved with conventional U-Pb zircon dating techniques [14,15], zircons from polymetamorphic rocks commonly exhibit extremely complex growth and compositional relationships that are impossible to resolve with these methods. Recent developments at the Australian National University in high resolution ion microprobe instrumentation (SHRIMP) and analytical techniques have provided a means of overcoming many of the limitations inherent in conventional U-Pb zircon analysis. In this abstract, ion microprobe U-Pb results for zircons from three Uivak I gneisses and one specimen of Uivak II gneiss, from the Saglek-Hebron area of Northern Labrador are reported. These results are compared with interpretations based on published conventional U-Pb zircon results and with conclusions about crustal evolution in the NAC derived from Rb-Sr, Sm-Nd and Pb-Pb isotopic studies.

*Geological Background.* Detailed accounts of the geology of the Archaean gneiss complex in Northern Labrador are given in [8,16,17]. The Uivak gneisses, a composite group of orthogneisses, have been subdivided on the basis of field relationships into two groups. The most abundant of these, the Uivak I suite, are dominantly fine-to-medium grained tonalitic and trondhjemitic gneisses with layering on scales ranging in width up to c. 100 cm. The layering is commonly accentuated by several generations of concordant to slightly discordant Na- and K-feldspar-rich pegmatite veins. Mineral assemblages in the Uivak I gneisses include quartz-oligoclase-microcline (or orthoclase) biotite  $\pm$  hornblende. Accessory phases are dominated by sphene, apatite and zircon. The Uivak II suite of K-feldspar-bearing augen gneisses contains a higher modal content of biotite and iron-rich amphibole. The presence of layered Uivak I gneiss xenoliths in outcrops of relatively undeformed Uivak II gneiss demonstrates that the deformations responsible for the formation of composite layering in the Uivak I gneisses occurred prior to the emplacement of the K-feldspar megacrystic granitic and granodioritic protoliths of the augen gneisses. Both members of the Uivak gneiss suite contain inclusions of older supracrustal rocks (the Nulliak assemblage). These range in size up to c. 2 x 0.25 km and are dominated by amphibolites of ultrabasic and basic composition as well as banded iron formation.

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Previous geochronological studies of the Uivak gneisses have yielded Rb-Sr, Sm-Nd, and Pb-Pb ages of  $3714 \pm 400/-291$  Ma ( $I_{Sr}$  0.69938  $-334/+252$ ;  $\epsilon_{Sr}$   $-11.8 \pm 18$ ),  $3612 \pm 379/-279$  Ma ( $I_{Nd}$  0.50722  $-22/+16$ ;  $\epsilon_{Nd}$   $+1.7$ ) and  $3572 \pm 318$  Ma ( $^{238}U/^{204}Pb$   $\mu_1 = 7.79$ ), respectively [10]. The isochron data for all 3 methods are relatively poorly correlated and hence exhibit large uncertainties; reflecting the combined influence of source heterogeneity as well as open system behaviour during later metamorphic disturbances. If Nulliak assemblage mafic rocks are the source of the tonalitic protoliths of the Uivak I gneisses, then it is valid to regress them with Sm-Nd data for the Uivak I gneisses, which gives  $3665 \pm 104$  Ma ( $I_{Nd}$  0.50719  $\mp 8$ ;  $\epsilon_{Nd}$   $+2.48 \mp 1.10$ ). The positive  $\epsilon_{Nd}$  value indicates that the Uivak I gneisses were derived from depleted mantle. The involvement of depleted mantle beneath the NAC in the formation of the precursors of the early Archaean tonalites is interpreted as reflecting the formation of still older continental crust.

Sr isotopic data for eleven large specimens of Uivak II gneiss yields a poorly fitted isochron (MSWD=248) with slope equivalent to an age of  $3412 \pm 158$  Ma ( $I_{Sr} = 0.69982 \mp 250$ ). However, whole-rock Pb isotopic results are better correlated (MSWD=13.4) and yield an isochron equivalent to an age of  $3703 \pm 293$  Ma, with a calculated  $\mu_1$  value for their source region of 7.65.

*Uivak Gneiss Zircon Results.* (1) Conventional Data. Previously published conventional U-Pb zircon analyses on both unsorted multi-grain samples of Uivak I and Uivak II gneiss [5] and zircon size-fractions from a single sample of Uivak II gneiss [18] are all highly discordant with  $^{207}Pb/^{206}Pb$  ages ranging from 2690 to 3485 Ma. With the exception of three of the Uivak I zircon analyses, which show the effect of recent Pb loss, seven multi-grain samples of Uivak I gneiss zircons are moderately well correlated (MSWD=34) and define a linear trend with a lower intercept of  $2600 \pm 67/-80$  Ma and an upper intercept with Concordia of  $4377 \pm 234/-297$  Ma. Data for the Uivak II gneiss zircon size-fractions, when combined with two multi-grain analyses, are extremely well correlated (MSWD=2.0) and define a chord with a lower intercept of  $2540 \pm 26/-28$  Ma (within error of the Uivak I gneiss zircon result) and an upper intercept of  $3950 \pm 88$  Ma. When regressed alone, the size-fractioned sample yields a Model 1 solution (MSWD=0.1) with significantly larger uncertainties in the upper intercept  $3760 \pm 387/-281$  Ma and a lower intercept of  $2490 \pm 95/-137$  Ma. Although these arrays may be the result of a single episode of Pb loss and have genuine age significance, the interpretation of the older Concordia ages must remain equivocal in view of the polymetamorphic history of the gneiss complex.

(2) Ion Microprobe Results. In an attempt to clarify interpretation of the conventional zircon results, zircon populations for three Uivak I gneisses and one Uivak II gneiss were analysed using the ion microprobe (SHRIMP) at the Australian National University [19,20]. Under routine operating conditions, the mass resolution of  $\leq 7000$  is sufficient to separate all significant spectral interferences, which obviates the necessity for peak stripping [cf. 21]. Methods have also been developed for determining Pb/U and Th/U ratios of unknown zircons to a precision of c. 3% [22]. The majority of the analyses are based on the means of three analyses achieved over a period of c. 45 minutes on the same spot. In most cases, the precision of the mean  $^{207}Pb/^{206}Pb$  per spot was typically between 0.5 and 0.8% (1 $\sigma$ ), limited principally by ion-counting statistics.

In terms of morphology, the Uivak I gneiss zircons generally exhibit a rounded core up to 100  $\mu m$  in diameter surrounded by euhedral to subhedral rims that commonly display well preserved growth zones. In contrast to the highly discordant conventional analytical results for the Uivak I zircons, the majority of the cores plot within error of Concordia or define a number

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of slightly discordant populations with ages between c. 3600 and c. 3920 Ma (Fig. 1). This range is similar to that shown by the  $^{207}\text{Pb}/^{206}\text{Pb}$  ages (Fig. 2).

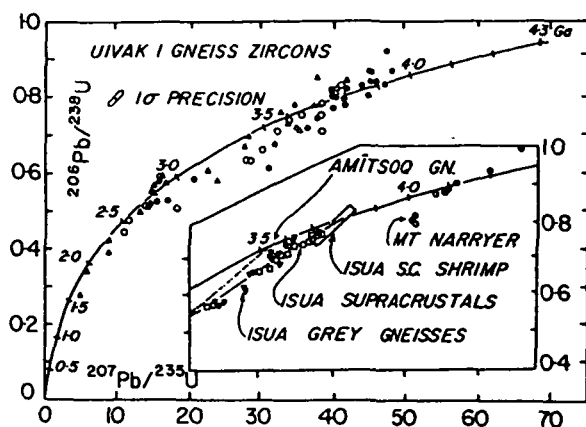


Fig. 1: Concordia diagram showing SHRIMP zircon analyses for three specimens of Uivak I gneiss; KC-78-633D open circles, and KC-78-620A triangles. The 1 $\sigma$  precision limits represent 3% uncertainty in determining Pb/U. The error box shown was calculated for the "old" zircon component. For comparison data are given in the insert from Mt. Narryer [23], the Amitsq gneisses [2, 24], the Isua grey gneisses [25] and the Isua supracrustal sequence [27, 23].

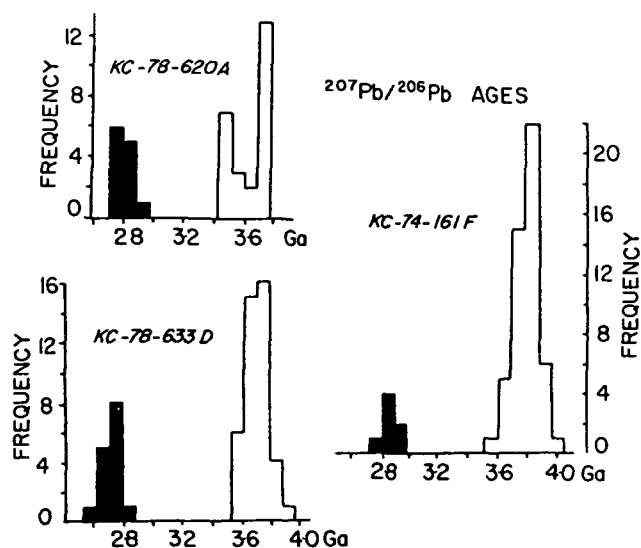


Fig. 2: Histogram showing the range of  $^{207}\text{Pb}/^{206}\text{Pb}$  ages for near concordant Uivak I gneiss zircons. The cores are shown without ornamentation, and the rims with solid ornamentation. The uncertainty in determining  $^{207}\text{Pb}/^{206}\text{Pb}$  ratios is generally better than 0.8%. This is significantly less than the age variation level.

Although most of the old cores contained relatively low amounts of Pb (18–100 ppm), U (less than 100 ppm), and Th (less than 50 ppm), a few cores were observed which contained substantially higher amounts of these elements; up to 1320 ppm Pb, 1380 ppm U, and 212 ppm Th. In a number of grains, cores showed considerable variation in  $^{207}\text{Pb}/^{206}\text{Pb}$  during sputtering, presumably reflecting compositional variation on a scale of c. 25  $\mu\text{m}$  x 3  $\mu\text{m}$ . For example, in one grain from KC-74-161F  $^{207}\text{Pb}/^{206}\text{Pb}$  ages varied from 4008 Ma through 3954 Ma to 3914 Ma. In Figure 1, analyses which lie above Concordia could reflect uncertainty in determining the Pb/U ratio. Alternatively, the reverse discordance may be the result of early gain of radiogenic Pb or loss of U. This feature has also been observed in microprobe analyses of zircons from elsewhere, e.g. Mt Narryer, Figure 1 [23].

The majority of zircon analyses with  $^{207}\text{Pb}/^{206}\text{Pb}$  ages between 3500 and 3000 Ma are relatively discordant (Fig. 1). This is interpreted to be the

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result of Pb-loss at between c. 3600 - 3800 Ma and c. 2800 Ma, as well as recent Pb-loss. In general, the higher U grains are the most discordant [cf. 24,25]. Rims of zircon typically have significantly higher contents of Pb, U and Th than are generally present in the older core regions. Most plot within error of Concordia at c. 2700 to 2900 Ma. This result is in excellent agreement with previously published estimates for a period of late Archaean regional metamorphism in the NAC [3,8,11,26].

Data for three zircons from the Uivak II gneiss plot close to Concordia on a chord of virtually constant  $^{207}\text{Pb}/^{206}\text{Pb}$  ratio which is equivalent to an age of c. 3350 Ma. A single rim analysis lies within error of Concordia at c. 3000 Ma.

*Discussion and Conclusions.* Analytical results for zircon cores in the three specimens of Uivak I gneiss described in this paper are significantly older than published conventional U-Pb analyses of zircons from the Amîtsoq grey gneisses, viz.  $3595 \pm 50$  Ma [2] and  $3575 \pm 50$  Ma [24]. Many of the analyses are also older than the c. 3700 Ma zircon population in the Isua grey gneisses [25]. These data are plotted for comparison in Figure 1. The c. 3600 Ma age for the Amîtsoq gneiss zircons is interpreted therefore as a metamorphic age [24,25]. In Figure 1 it is clear that most of the Uivak I gneiss zircon data are broadly within error of conventional [27] as well as SHRIMP [23] results for time of crystallization of zircons from the Isua supracrustal sequence, viz.  $3813 \pm 6/-4$  Ma. From this, it follows that zircon cores in the Uivak I gneisses may be xenocrystic, representing compositions inherited from a c. 3800 Ma old source which melted to form the plutonic precursors of the Uivak I gneisses. Alternatively, the zircons may have crystallized c. 3800 Ma ago in the plutonic protoliths of the Uivak I gneisses. The low Pb, U and Th content of most of the Uivak I gneiss zircon cores either reflects the relatively basic character of the crustal precursors of the gneisses, or it may be typical of the compositional range of zircon crystallizing from melts of tonalitic and trondhjemitic composition.

The dispersion observed in zircon cores from the Uivak I gneisses between c. 3800 Ma and 3600 Ma is interpreted to be the result of variation in the degree of loss of radiogenic Pb in response to younger periods of metamorphism, together with the effect of recrystallization during such thermal events.

Several virtually concordant low and high U grains have  $^{207}\text{Pb}/^{206}\text{Pb}$  ages that are in excess of 3800 Ma, ranging up to c. 3920 Ma. These provide unequivocal evidence of pre-"Isua" inherited xenocrystic components in the zircons. It is concluded from the range of Pb and U in these xenocrysts that basic as well as LIL element enriched (acid) crustal compositions were present in the source of the protoliths of the Uivak I gneisses. This supports interpretations based on Sm-Nd studies of the Uivak gneisses [10] that they were derived from a depleted source from which earlier crust has been extracted. As none of the ion microprobe data yield ages in excess of 4000 Ma, little geological significance can be ascribed to the value of the upper Concordia intercept defined by the conventional Uivak I gneiss zircon data.

The current interpretation based on field and geochemical evidence is that the protoliths of the Uivak II gneisses were derived by partial melting of pre-existing Uivak I gneiss - Nulliak assemblage crustal components. The involvement of such old source components is supported by previously discussed Pb-Pb whole rock isotopic data and also by conventional U-Pb zircon results. However, the whole rock Sr isotopic data and the zircon ion microprobe data currently available yield significantly younger ages (c. 3350 Ma). This is interpreted to indicate that the megacrystic granite protoliths of the Uivak II gneisses were formed and metamorphosed c. 3350 - 3400 Ma ago, and were contaminated with old crustal unradiogenic Pb [cf. 11]. A second

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scenario is that the Sr and SHRIMP results currently available were reset by metamorphic recrystallization processes and therefore they date the time of fabric development in the gneisses. Failure to identify an old component, consistent with the interpretation of the conventional zircon data, is interpreted to be a reflection of the small number of analyses currently available.

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THE TRANSITION FROM AN ARCHEAN GRANITE-GREENSTONE TERRANE INTO A CHARNOCKITE TERRANE IN SOUTHERN INDIA; Kent C. Condie and Philip Allen, Dept. of Geoscience, New Mexico Institute of Mining and Technology, Socorro, NM 87801

In southern India, it is possible to study the transition from an Archean granite-greenstone terrane (the Karnataka province) into high grade charnockites (Fig. 1). The transition occurs over an outcrop width of 20-35 km and appears to represent burial depths ranging from 15 to 20 km (1). Field and geochemical studies indicate that the charnockites have developed at the expense of tonalites, granites, and greenstones (1, 2, 3, 4). South of the transition zone, geobarometer studies indicate burial depths of 7-9 kb (5).

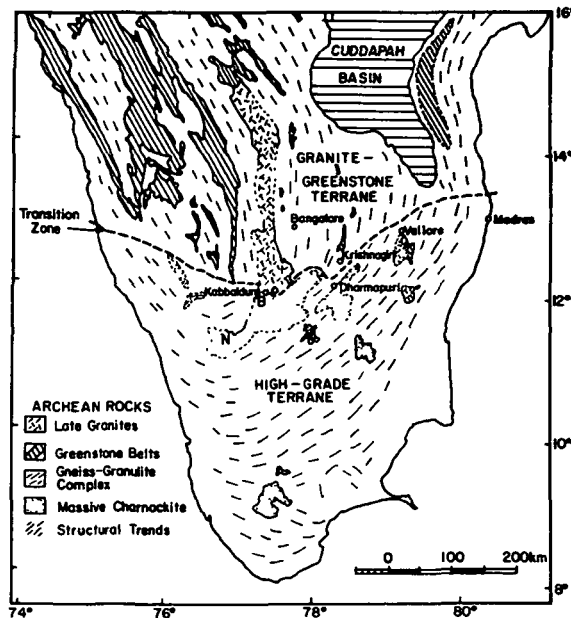


Figure 1. Geologic map of southern India showing the transition zone.

On most chemical variation diagrams, the charnockites, define igneous-like trends with a sparsity of intermediate compositions ( $\text{SiO}_2 = 55\text{-}65\%$ ) and a great majority of tonalitic compositions. As with other granulite-facies terranes, the Indian charnockites exhibit significant depletions in Rb, Cs, Th, and U (4). However, a well-defined correlation of degree of depletion and metamorphic grade is not apparent. Many of the low-grade charnockites exhibit as great of depletion as high-grade charnockites. Unlike most other high-pressure charnockites, however, Indian charnockites do not exhibit

depletion in K, and Rb depletion does not reach the 1-2 ppm level as observed in Lewisian granulites. Indian tonalites and charnockites do not exhibit significant differences in the contents of Ba, Sr, REE, high field strength elements or transition metals. Depletions in Rb, Cs, Th, and U appear to have occurred during the passage of a CO<sub>2</sub>-rich fluid phase.

Tonalitic and granitic charnockites (like their protoliths) have light-REE enriched patterns and generally exhibit an inverse correlation between SiO<sub>2</sub> and total REE content. Eu anomalies range from slightly negative to strongly positive, with Eu/Eu\* exhibiting an inverse correlation to REE content.

With exception of Rb, Cs, U, and Th, the major and trace element distributions in Indian charnockites reflect the composition of their protoliths. Geochemical modelling clearly indicates that the tonalitic charnockite protoliths (TCP) have been produced by partial melting of a mafic source rich in hornblende and/or garnet or by fractional crystallization of a wet basaltic magma (1, 4). A mafic source must be enriched in incompatible elements relative to N-MORB or TH1 (the major Archean basalt type) and depleted relative to continental rift basalts (Fig. 2). Such a source is similar in composition to TH2 (enriched Archean basalts). Archean tonalites, in general, demand the existence of an enriched mafic source and thus require a substantial volume of enriched mantle by the late Archean.

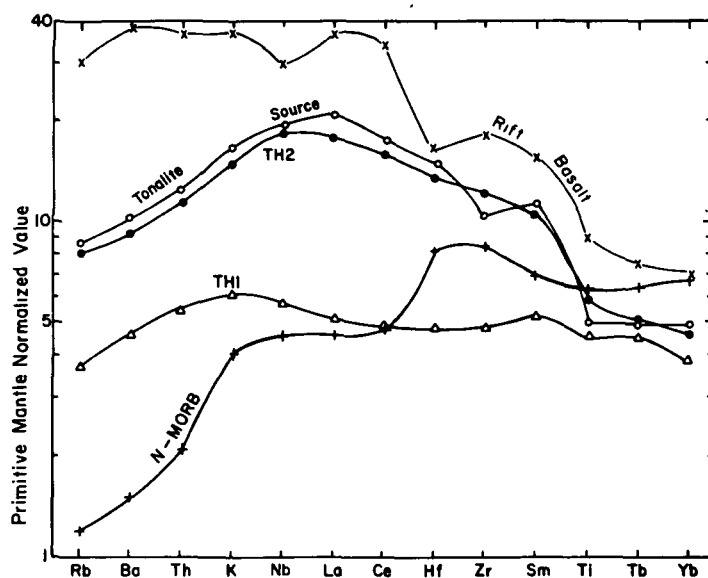


Figure 2. Primitive mantle normalized incompatible element distributions in the Indian TCP source compared to other mafic compositions (modified from 1). Rb and Th contents of TCP from Indian tonalites.

Most Indian charnockites cannot represent the residue left after extraction of granitic magma as indicated by the contrasting incompatible element contents (Fig. 3B). One sample from the transition zone (NC1), however, does match the calculated residue. This suggests that although the charnockite terrane as a whole cannot be considered as residue, portions of the terrane in the transition zone may represent residue.

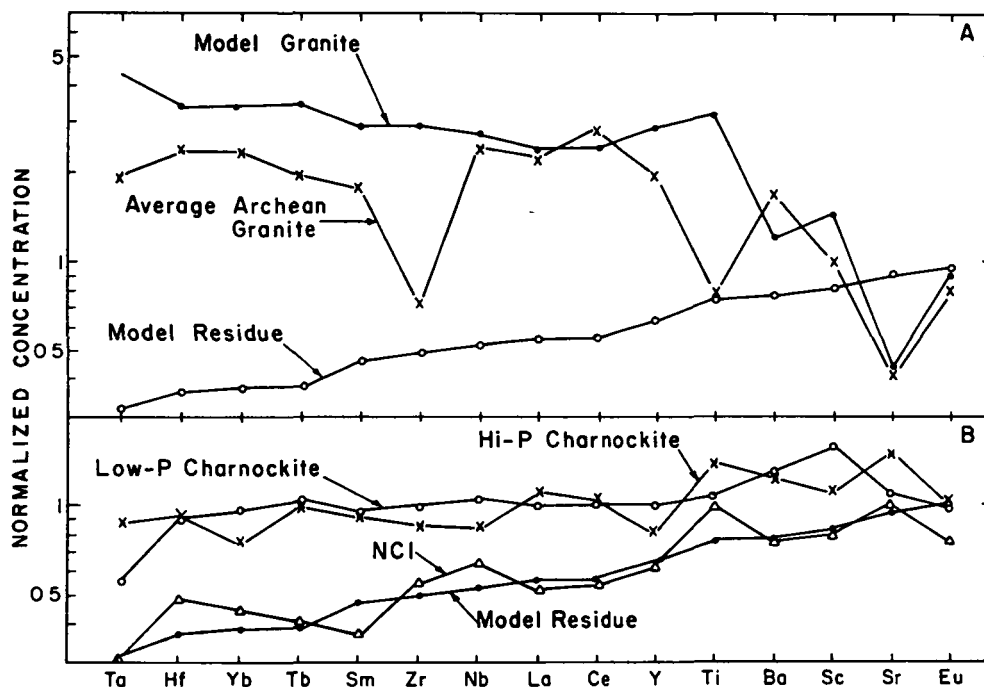


Figure 3, Tonalite-normalized element distributions of relatively incompatible elements in granite and charnockites. Model granite and residue are produced by 20% batch melting of tonalite. Mode and melt fractions given in (4).

Although field and geochemical data clearly indicate that some granites in the transition zone are of metasomatic origin (1), others may be partial melts and still others, involve both processes (6). Field relationships along the transition zone strongly suggest that a fluid phase relatively rich in  $\text{CO}_2$  purged  $\text{H}_2\text{O}$  from the lower crust and concentrated it in a relatively narrow region at mid-crustal levels where partial melting produced migmatites with granite leucosomes (1, 3, 6). In some areas along the transition zone, major plutons may have formed by such a mechanism. Thus, granite formation in the Archean crust of India is closely related to charnockitization and is localized, for the most part, at intermediate crustal levels (20-25 km) along a metasomatic front. Deep crustal charnockites represent tonalites in which large amounts of Rb, Cs, Th, and U have been

removed by fluids with variable, but on the whole, high  $\text{CO}_2/\text{H}_2\text{O}$  ratios. High  $\text{CO}_2/\text{H}_2\text{O}$  ratios in fluids also raise solidus temperatures and thus prevent partial melting of the lower crust. If this model is correct, it is the middle and not the lower crust where we should look for the residues left after granite extraction.

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AVERAGE SEDIMENTARY ROCK RARE EARTH ELEMENT PATTERNS AND CRUSTAL EVOLUTION: SOME OBSERVATIONS AND IMPLICATIONS FROM THE 3800 Ma ISUA SUPRACRUSTAL BELT, WEST GREENLAND. R.F. DYMEK, Dept. of Geol. Sci., Harvard Univ., Cambridge, MA 02138; J.L. BOAK, Arco, P.O. Box 360, Anchorage, AK 99510; L.P. GROMET, Dept. of Geol. Sci., Brown Univ., Providence, RI 02912

INTRODUCTION: The hypotheses that rare earth element (REE) abundances and patterns in clastic sedimentary rocks trace the unidirectional chemical evolution of the upper continental crust, and that the "average" REE pattern of Archaean sediments is fundamentally different from that observed in post-Archaean sediments, have been advanced in recent years (e.g. [1]). If correct, such conclusions indicate substantial differences in the average composition of Archaean crust, and place important constraints on growth vs. no-growth models of continental development (e.g. [2,3]).

These hypotheses rely heavily on three very key assumptions: (a) REE experience no relative fractionation during weathering, erosion, deposition and diagenesis accompanying the transformation of igneous rock into sediment; (b) REE in sediments provide a broad average of available source areas at the time of sedimentation; and (c) sampled units are representative of sediment deposited in the area at the time of formation. Moreover, in the case of meta-sediments, with which one is commonly faced in ancient terranes, it is also assumed that metamorphism (at least through medium grades prior to the onset of melting) does not perturb whole-rock REE patterns. Collectively, these assumptions outline the rationale for linking sediment REE patterns to those in igneous rocks. However, none of these assumptions have been tested extensively, although broad similarities among REE patterns in Phanerozoic sediments (cf. [4]) seem to support the viewpoint that averaging of source area REE does in fact occur, given the diversity of REE patterns in crustal igneous rocks.

We are, however, quite concerned with point (c), i.e., whether published REE patterns on Archaean sediments are representative of Archaean sediments in general, whether they reflect accidents of preservation, or whether there is an inherent bias in the data base, albeit unintentional. For example, anyone familiar with studies of Archaean geology will recognize the nearly complete absence of data on sediments from high-grade terrains, whereas most data are for graywackes from low-grade greenstone belts, which, as pointed out by Pettijohn [5], bear a strong resemblance to Phanerozoic eugeosynclinal sediment suites.

In this report, we present REE data on a set of clastic metasediments from the 3800 Ma Isua Supracrustal belt, West Greenland. Each of two units from the same sedimentary sequence has a distinctive REE pattern, but the average of these rocks bears a very strong resemblance to the REE pattern for the North American Shale Composite (NASC), and departs considerably from previous estimates of REE patterns in Archaean sediments. We regard the possibility that the source area for the Isua sediments discussed here resembled that of the NASC as highly unlikely. However, REE patterns like that in the NASC may be produced by sedimentary recycling of material yielding patterns such as are found at Isua.

GEOLOGICAL SETTING: The Isua supracrustals are located ~140 km northeast of Godthåb, central West Greenland, where they crop out in an arcuate belt surrounded and locally intruded by ca. 3700 Ma Amitsoq orthogneiss. Ages on various supracrustal units are in the 3700-3800 Ma range, with the most precise determination being  $3769 \pm 11$  Ma by U-Pb methods on single zircons [6]. Of particular significance to the present study is a Sm-Nd whole rock isochron age of  $3770 \pm 130$  Ma [7] on leucoamphibolite ("garbenschiefer" formation), which establishes a minimum age of deposition for the protolith of the

metasediments, as this unit is apparently intrusive into the section.

Several descriptions of Isua geology have been published [10,11], which outlined the presence of metavolcanic and clastic and chemical metasedimentary units. More recent detailed mapping [12] has revealed a coherent stratigraphy. The main supracrustal group - Sequence A - crops out along the entire length of the belt, and contains various amphibolites, quartz-rich chemical metasediment (including carbonate, silicate and magnetite ironstone), calc-silicate gneiss, felsic muscovite-biotite gneiss, and minor garnet-biotite schist. A tectonically separate group - Sequence B - crops out only in the eastern part of the belt, and consists of a lower unit of felsic muscovite-biotite gneiss (MBG) and an upper unit of predominantly garnet-biotite schist (GBS). The contact between the lower and upper units of Sequence B is gradational. MBG have been referred to as "pelitic metagraywackes", whereas GBS represents a series of "ferruginous shales" comprising pelitic to semipelitic to mafic types.

The relationship between Sequences A and B is unclear: Sequence B may be older than or correlate to the lower part of Sequence A, but substantial differences in chemical composition suggest no direct relationship. In this report, we discuss only samples from Sequence B.

**RESULTS:** REE were analyzed by isotope dilution mass spectrometry; chondrite-normalized data for 12 samples are illustrated on Figure 1.

MBG have REE patterns that are enriched and moderately fractionated with respect to chondrites ( $Ce_N/Yb_N=6.8-8.0$ ), with highly variable negative Eu anomalies ( $Eu/Eu^*=0.45-0.96$ ) and a slight flattening in the heavy REE. The similarity in REE pattern shape for these samples, which span a wide range of bulk composition (e.g.,  $SiO_2=58-76$  wt %), suggests that a single component or a relatively constant mixture of components dominates the REE characteristics of MBG. Some Archaean felsic igneous rocks have REE patterns not unlike the MBG [13].

GBS have REE patterns that are less enriched and less fractionated than MBG, and two pattern shapes are discernible. The four samples with lowest REE abundances, which represent pelitic and semipelitic rocks (garn + bio ± musc ± stl), have fractionated light REE, small Eu anomalies of variable sign, and a slope reversal for the heavy REE (i.e.,  $Gd_N < Yb_N$ ). Here again, the similarity in pattern shape suggests a single REE component or mixture of components, although we are unaware of any igneous rocks that have REE pattern shapes like these four GBS samples.

The fractionated light REE suggest a contribution from material not unlike that of the MBG, with which the GBS are locally interlayered. High Cr and Ni contents in GBS (up to 850 and 350 ppm respectively) may indicate a contribution from mafic material. Mixing of REE derived from basaltic rocks (low abundances, unfractionated patterns) with REE derived from felsic rocks (high abundances, fractionated patterns) could explain the lowered light REE abundances in GBS, but we are unable to provide a completely satisfactory explanation for the slope reversal in the heavy REE.

The fifth GBS sample (28-6A), which represents a mafic metasediment (garn + bio + hbl), has relatively unfractionated light REE, a small positive Eu anomaly and only slightly fractionated heavy REE. The REE pattern shape for this sample strongly resembles that found in Isua amphibolites [8,14], except that it is enriched by a factor of two. This suggests that the REE in this sample were derived almost exclusively from a mafic source.

**DISCUSSION:** Although the REE patterns for MBG and GBS are clearly different (Figure 1), the fact that these units are interlayered in the field indicates that they were deposited penecontemporaneously and sampled a diverse suite of crustal materials that were in existence at the time of deposition.

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Hence, their REE provide some type of estimate of source area characteristics, but how one constructs an average REE pattern and what that average means, are problematic. Moreover, there is no guarantee that mixing proportions of REE correspond in a direct way to volume proportions of crustal rocks.

The average REE patterns for MBG and GBS are shown in Figure 2. Both convey the general pattern shape of the individual samples in each unit (except for GBS 28-2A as noted above; however, it would be inappropriate to exclude this from the average; were this analysis in fact excluded, it would not change the shape of the average GBS pattern in any significant way). Two average REE patterns for Sequence B are also indicated in Figure 2. The first (left) is simply the arithmetic mean of the twelve analyzed samples, and is shown for illustrative purposes only. The second (right) was calculated by weighting each unit in proportion to its abundance in the field (3 parts MBG: 2 parts GBS). Important features of the Sequence B average meta-sediment pattern are the enriched and fractionated light REE and the substantial negative Eu anomaly. Also illustrated in Figure 2 (right) are REE patterns for the North American Shale Composite (NASC: [9]) and for average Archaean sediment (AAS: [1]). The similarity in REE pattern shape between the Isua average and NASC is evident, as is the difference between AAS and NASC. These features are emphasized further in Figure 3, where both the Isua Sequence B average and AAS are normalized to the NASC. The relatively flat unfractionated pattern for the Isua average is particularly noteworthy.

**CONCLUSIONS:** The results reported here lead to the following tentative conclusions. (1) The REE patterns for Isua Seq. B MBG indicate the existence of crustal materials with fractionated REE and negative Eu anomalies at 3800 MA. Processes such as feldspar fractionation in shallow level magma chambers or intracrustal partial melting may have been important in the development of the sediment source rocks. (2) The average Seq. B REE pattern resembles that of the NASC. The methods by which average REE patterns for sediments are determined, and what the significance of such averages is, require further evaluation. (3) If the Seq. B average is truly representative of its crustal sources, then this early crust could have been extensively differentiated. In this regard, a proper understanding of the NASC pattern, and its relationship to post-Archaean crustal REE reservoirs, is essential. (4) The Isua results may represent a "local" effect. Additional study of Archaean sediment REE characteristics, especially those in high-grade terrains, are warranted.

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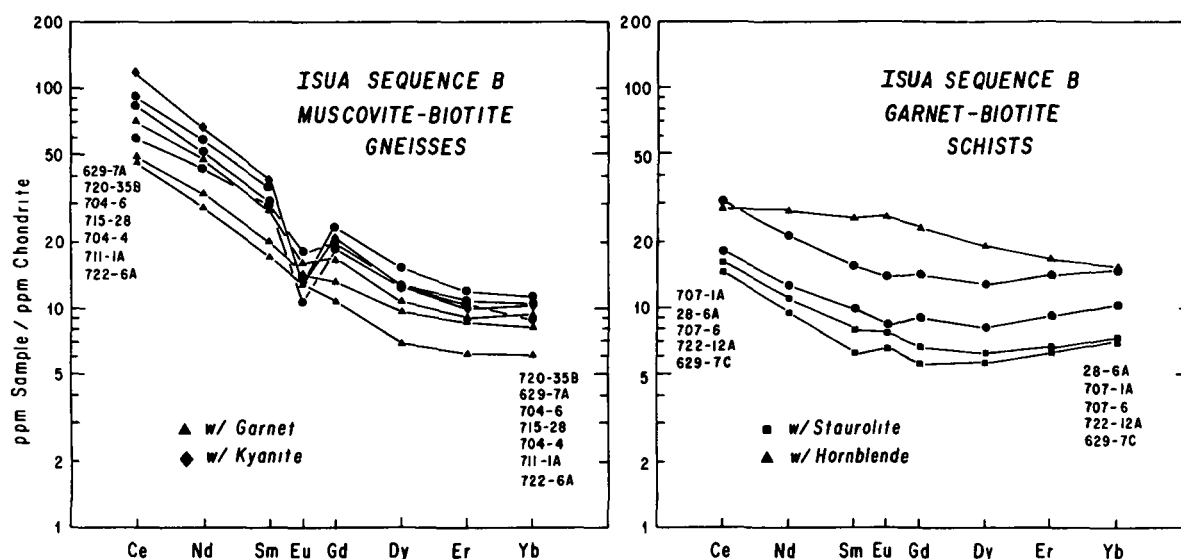


Figure 1. REE patterns for Isua Seq. B metasediments. The data for 629-7C are revised from those reported previously [8], based on reanalysis using HF bomb dissolution. Reanalysis of selected other samples confirms original patterns.

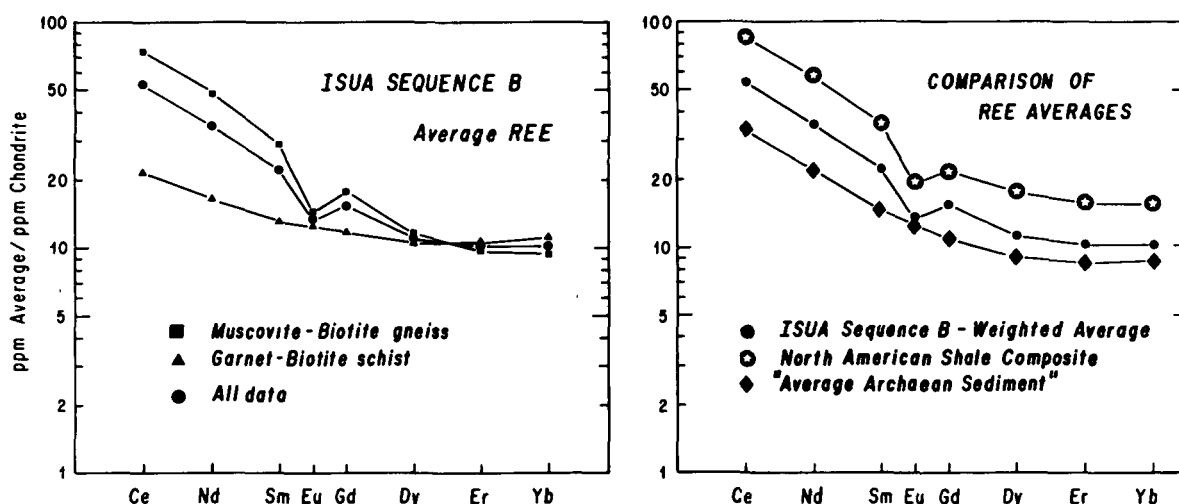


Figure 2. (left) Average REE patterns for data shown in Fig. 1; "all data" is the arithmetic mean of the 12 analyzed samples. (right) Isua Seq. B weighted average (3 parts MBG: 2 parts GBS) compared to REE patterns for NASC [9] and for average Archaean sediment (AAS) [1].

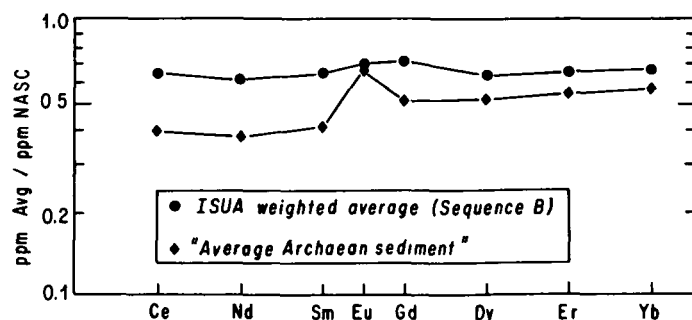


Figure 3. Comparison of NASC-normalized REE patterns for Isua Sequence B weighted average and "average" Archaean sediment (AAS). Note relatively "unfractionated" pattern for the Isua average, whereas AAS shows light REE "depletion", positive "Eu anomaly", and low-overall REE concentrations.



STRUCTURAL VARIATION OF PROTEROZOIC DIKES IN THE  
CENTRAL SUPERIOR PROVINCE- A POSSIBLE REFLECTION OF  
POST-ARCHEAN SHIELD DEFORMATION, R. E. ERNST and H. C. HALLS,  
Dept. of Geology, University of Toronto, Toronto, Canada, M5S  
1A1

### Introduction

Geodynamic modelling of shield areas requires a 3-D understanding of the geology. In lieu of deep drilling and geophysical work, the third dimension is revealed wherever the crust is tilted to expose a vertical section at surface. However, the identification of surface exposure with changing structural depth is generally difficult in shield areas owing to their typically complex deformation. In areas where the structural and metamorphic patterns provide inconclusive results, data from post-orogenic diabase dike swarms may prove useful.

Undeformed mafic dike swarms of simple geometry cross-cut most shields. It has been suggested that their structural, paleomagnetic, and chemical characteristics can provide estimates of variation in exposure-depth throughout a shield terrain [1,2,3].

Preliminary work is reported below on the use of two structural parameters, dike dip and thickness, as possible depth-of-exposure indicators in the Central Superior Province.

### Dike Dips

The dike dip data for the Central Superior Province divide into geographic domains of like dip (Figure 1). Domains of non-vertical dip may originate either because the dikes were intruded in a near vertical attitude and subsequently tilted along with the host terrain or because the contemporaneous stress regime or inherited mechanical anisotropy favoured non-vertical intrusion. While intrusion along an inclined plane almost certainly explains some of the scatter in our data, it appears that the dominant cause of regionally inclined dike-dips in the Central Superior Province is post intrusion deformation. This interpretation is based on the following observations: that, in general, dike orientation is unrelated to any obvious host rock grain, and more importantly, that the sense of dip in each domain is consistent with tilting of the host terrain as given by metamorphic, tectonic and paleomagnetic evidence (Figure 1, Table 1., [4,5])

### Dike Thickness

The 3-D form of dikes is largely unknown but it has been speculated that dikes are vertically localized bodies which taper and pinch out with depth [1,6]. Therefore, progressive changes in dike thickness across or along a swarm may reflect differences in depth-of-exposure. However, this effect may be overprinted or totally masked by true lateral variation in dike thickness, resulting from variations in the elasticity of the host rocks or in the characteristics of the rising magma.

Preliminary dike-thickness data are available for the Central Superior Province and these data also appear to reflect host-rock erosional level (Figure 2); the average thickness of 2.6 Ga. dikes varies inversely with host rock metamorphic grade. Average thicknesses of less than 9 meters are found in

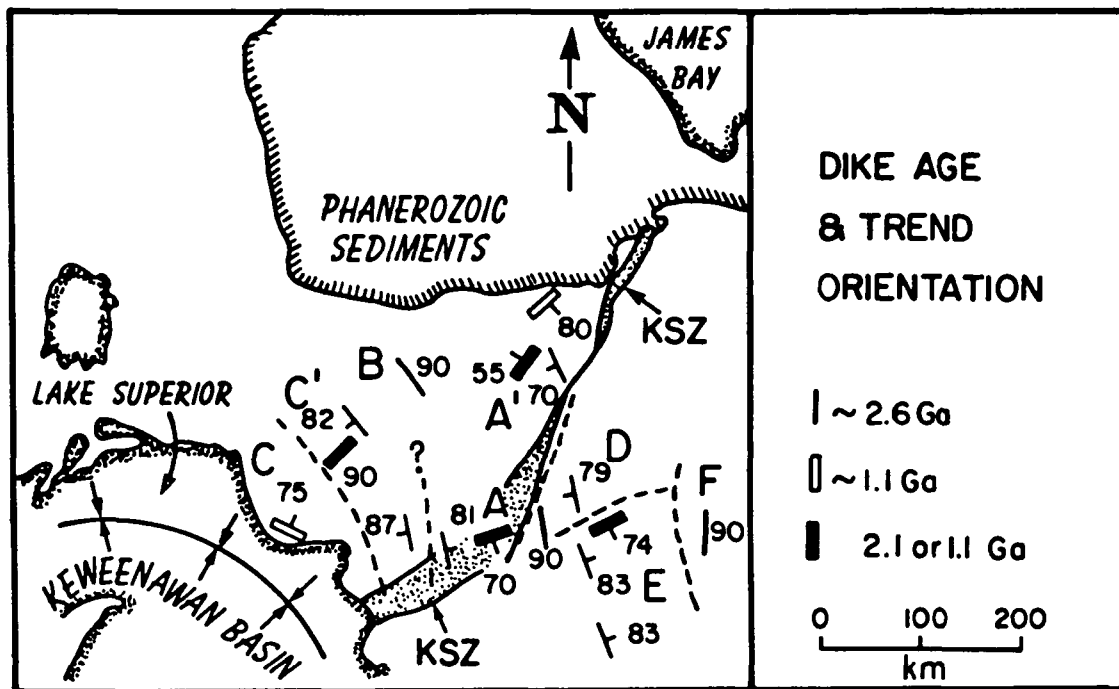


Figure 1. Dike Dip Domains in the Central Superior Province. Descriptions of each domain given in Table 1. Details of dip data given in references [3,5]. Number associated with each dike symbol gives the dip in the direction indicated by the dash. No dash indicates vertical dip. Underlined values based on 5-23 individual dike measurements, and the others 1-4 measurements. The average standard deviation for each determination is 7 degrees.

Table 1 DIKE DIP DOMAINS

Domain	Interpretation (according to model of post-intrusive deformation)
A	Block tilting uplift around NNW hinge line with detachment along the eastern margin of the Kapuskasing Structural Zone (KSZ) [8].
A'	Sparse data; non-vertical dips probably reflect block tilting associated with the KSZ
B	Outside zone of influence of KSZ
C	Associated with sagging of volcanic-laden crust of the Keweenaw Basin [9,10]
C'	These dikes are possibly located on the distal side of a peripheral bulge associated with down warping of the Keweenaw Basin.
D	Implies horizontal to shallow eastward dipping crust-the regional metamorphic grade also decreases in an eastward direction ([11] and Figure 2).
E	Implies shallow westward tilt of crust- perhaps results from dip slip movement along some of the numerous NNW-SSE striking faults which traverse this area. Northern boundary with D coincides with location of the large Abitibi dike shown in Figure 2, suggesting local emplacement of the dike along a fault. In this regard, -it is interesting that although the dike appears to pinch out towards the SW, its extrapolated extension coincides with a KSZ boundary fault. We may speculate whether emplacement of this dike allowed detachment and differential movement of domains D and E together with renewed KSZ uplift at 1.1 Ga. or whether the fault was active earlier and simply reused by the dike at 1.1 Ga.
F	Outside zone of influence of KSZ

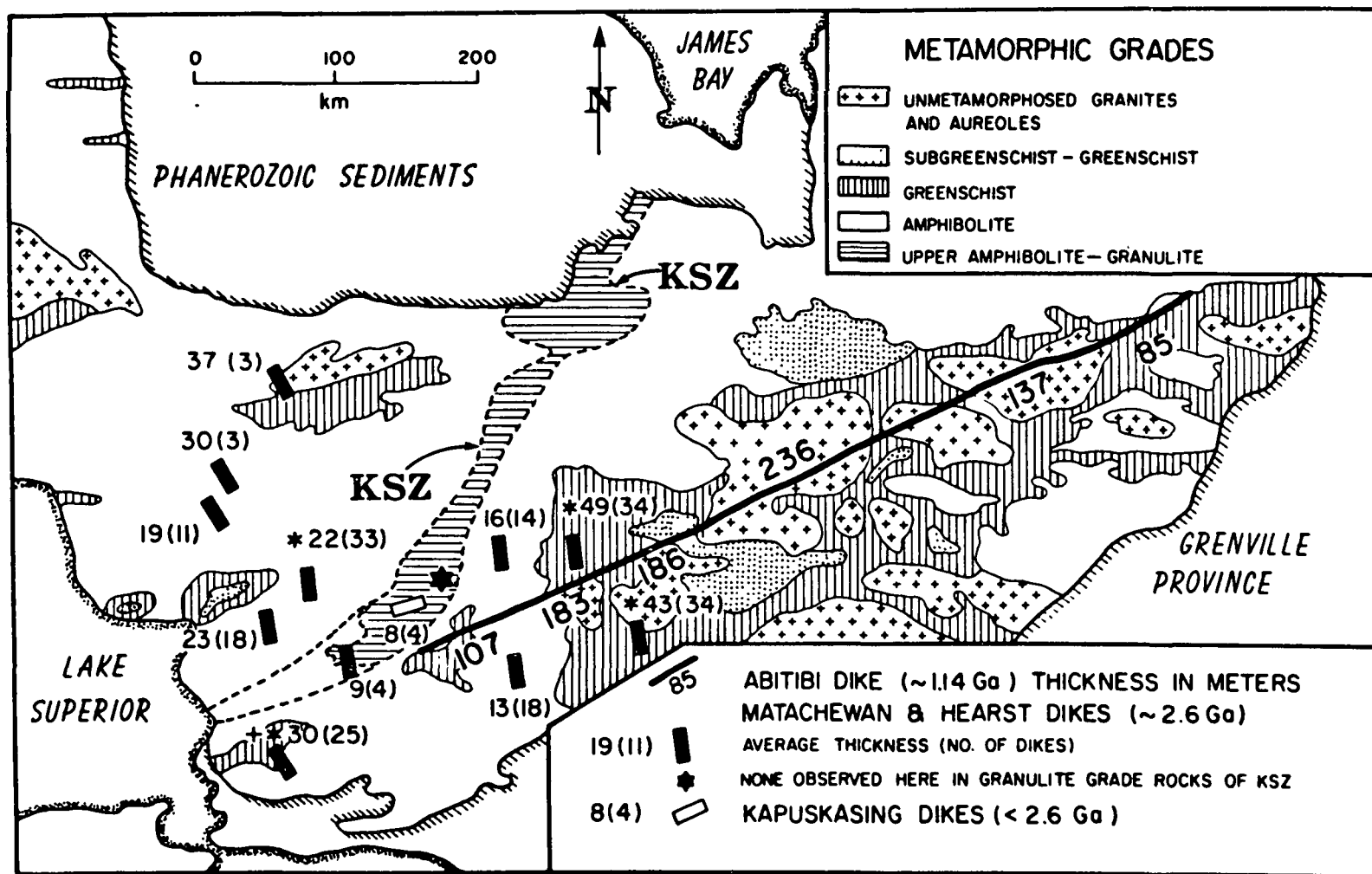


Figure 2. Regional Pattern of Dike Thickness in the Central Superior Province. The rectangular boxes give the generalized dike trend. Associated with each is a number giving the average dike thickness in meters and in parenthesis the number of dikes on which this is based. A star (\*) indicates measurements made using outcrop data from published maps [13, 14, 15, 16]; the remaining average thicknesses were determined in the field from roadside outcrops. The datum marked with a plus (+) is from possible post 2.6 Ga. dikes. The standard deviation of each determination is about 10 meters. No correlation was observed between dike thickness and host rock type. The generalized metamorphic pattern is from references 11 and 12. The age quoted for the Abitibi dike is a preliminary unpublished U/Pb determination (E. Nakamura & T. Krogh, pers. comm., 1983). Numbers beside the Abitibi dike give its local thickness in meters. Note the inverse correlation of dike thickness with host rock metamorphic grade for dikes of 2.6 and 1.14 Ga. age.

the upper amphibolite to granulite zones of the Kapuskasing Structural Zone (KSZ) and in particular, no 2.6 Ga. dikes are observed in the granulites [7]. These data contrast sharply with those from outside of the KSZ; average thicknesses of 16 to 30 meters occur within amphibolite grade rocks and values of 30 to 50 meters in host rocks with a greenschist to subgreenschist grade. Similarly, the Abitibi dike is thickest (240 m) where the surrounding rock is sub-greenschist to greenschist and thins towards areas where the regional host-rock grade is greenschist to amphibolite (Fig. 2).

These dike-thickness data can be modeled by a post 2.6 Ga. uplift of the southern part of the KSZ [8], and a post 1.14 Ga. upwarp of the east and west margins of the Abitibi Subprovince.

#### Conclusion

The above data demonstrate systematic variations in the dip and thickness of 2.6 and 1.14 Ga. dikes across the Central Superior Province and are tentatively interpreted to result from post intrusion deformation. Combination of these results with additional structural and paleomagnetic data from dikes of all ages may permit detailed mapping both spatially and temporally of crustal deformation in this part of the Canadian Shield.

Although dike dip and thickness data apparently reflect crustal exposure level as given by host rock metamorphic grade (ranging from subgreenschist to granulite), these post-orogenic dikes themselves are at most only weakly metamorphosed. This requires that regional isotherms dropped dramatically after the Kenoran orogeny (2.65 Ga.) and prior to emplacement of the earliest post-orogenic swarm (Matachewan-Hearst) at 2.6 Ga.

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GEOPHYSICAL CONSEQUENCES OF PHANEROZOIC AND ARCHEAN CRUSTAL EVOLUTION: EVIDENCE FROM CRUSTAL CROSS-SECTIONS; David M. Fountain, Dept. of Geology and Geophysics, Program for Crustal Studies, Univ. of Wyoming, Laramie, WY 82071

Geophysical properties of continental crust depend on the nature of crustal evolution. This is well illustrated by examination of two crustal cross-sections (1), the combined Ivrea-Verbano zone (IVZ) and Strona-Ceneri zone (SCZ) of northern Italy and the Pikwitonei granulite belt (PGB) and Cross Lake subprovince (CLS) of Manitoba. These two cross-sections are of particular interest because the IVZ and SCZ developed during Phanerozoic time whereas the PGB-CLS is an example of Archean crustal evolution. Consequently, each cross-section is geologically distinctive and, thus, exhibits very different geophysical properties such as density, seismic velocity, heat production, and magnetism.

Perhaps the best known cross-section of the crust is the IVZ-SCZ of northern Italy (Figure 1). Deeper crustal levels are represented by the granulite and upper amphibolite facies ultramafic, mafic and pelitic rocks of the IVZ that experienced peak metamorphic conditions of 9-11 kb and 700°-820°C during Caledonian time (2,3). Isotopic data indicate that these high-grade rocks resided at lower to middle crustal levels until Jurassic time. Amphibolite and greenschist facies pelitic schists and gneisses, intruded by post-metamorphic granitic plutons, dominate higher crustal levels exposed in the SCZ. Peak metamorphism in the SCZ was also Caledonian (2). The IVZ and SCZ are separated by a vertical mylonite zone, the Pogallo line, which Hodges and Fountain (4) interpret as a Jurassic low-angle normal fault rotated into its present position during Alpine deformation. The Pogallo line is but one manifestation of the rift event that created the Tethys Ocean.

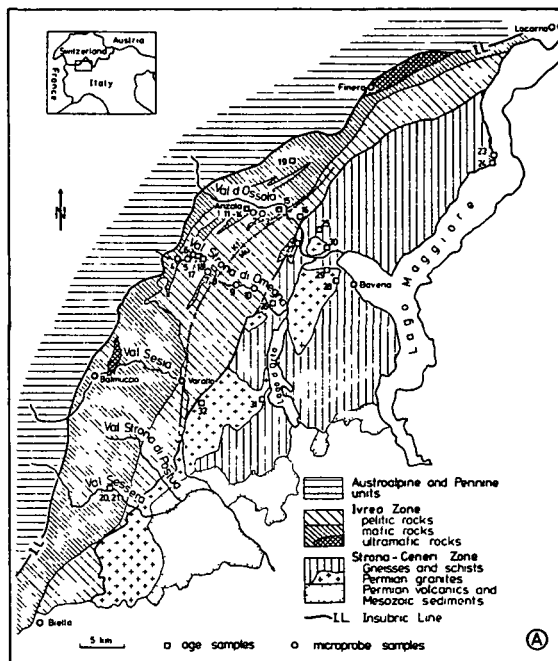


Figure 1. Sketch map of IVZ and SCZ from Hunziker and Zingg (2).

The early history of the IVZ and SCZ is somewhat enigmatic but was dominated by deposition of a thick pelitic package of sediments between 480 and 700 Ma (2). Deep burial and amphibolite facies metamorphism preceded

intrusion of mafic and ultramafic magmas and peak metamorphism during Caledonian time (2). Schmid (5) speculated that granulite facies pelitic gneisses were de-granitized during this event. Post-metamorphic intrusion of granites suggests Andean margin conditions prevailed later in the Paleozoic. Rifting, normal faulting and formation of a Tethyan trailing margin modified this Paleozoic crustal block during early Mesozoic time.

Geophysical properties of the IVZ and SCZ are a consequence of its Phanerozoic evolution. Seismic velocities (6) for mafic and ultramafic rocks are expectedly high (7.2–8.4 km/sec). Pelitic granulites also have high velocities (7.0–7.6 km/sec) because of high garnet and sillimanite content. Deep crustal levels are characterized by high P-wave velocities. In contrast, schists, gneisses and granites of the SCZ have P-wave velocities around 6.4–6.5 km/sec, causing a marked seismic distinction between the lower crust and higher levels. Investigations by Wasilewski and Fountain (7) indicate that magnetite is the magnetic phase in the granulites and that susceptibilities are high in mafic granulites and very low in higher level rocks of the SCZ. The deep crust represented by the IVZ, therefore, is also magnetically distinct and could cause long-wavelength magnetic anomalies if it resided in areas of normal geothermal gradients. Published heat production data (8) indicate that heat production is low in mafic granulites (0.8 HGU), moderately high in upper amphibolite pelitic gneisses and SCZ lithologies (5–6 HGU) and high for post-metamorphic granites (7 HGU). No data is available for pelitic granulites. These data suggest that there may be a significant stepwise decrease in heat production at deep crustal levels although analysis of high-grade pelitic rocks is necessary to confirm this.

Deep levels of Archean crust are exposed in the Pikwitonei granulite belt of Manitoba (Figure 2). About 80% of the PGB gneisses are silicic granulites (enderbites and charnockites) that surround much less abundant, discontinuous layers of granulite facies paragneisses, iron formation, and mafic and ultramafic rocks (9) that are similar to nearby lower grade greenstone belt lithologies. The CLS, representing shallower crustal levels, lies to the southeast of the PGB and consists of several greenstone belts surrounded by granites and granitic gneisses. Metamorphic grade in these belts tends to decrease to the southeast.

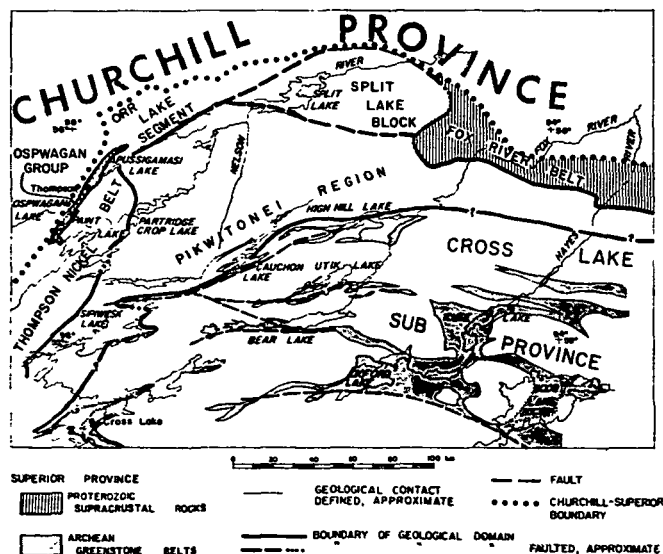


Figure 2. Sketch map of geological domains of northwestern Superior Province from Weber and Scoates (9).

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Geologic investigations in the PGB and CLS (9, 10) suggest that there is no fundamental difference between the two terrains other than variation of metamorphic grade and differences in volumetric abundances of certain supracrustal lithologies. Mapping by Hubregtse (10) in the Cross Lake area demonstrated that low-grade greenstone belts can be traced across the orthopyroxene isograd into the PGB, suggesting that the PGB and CLS share a common history. Isotopic data (11) suggest an approximate 2.8 Ga age of metamorphism for PGB enderbites. Ages of metaplutonic rocks of CLS are between 2.7-3.0 Ga (11).

The bimodal compositional nature of rocks of the PGB and CLS produce a distinctive geophysical character. Enderbites and granitic rocks surrounding supracrustal belts have similar mineralogy, low densities (mean density =  $2.68 \text{ g/cm}^3$ ) and thus, low seismic velocities (estimated at approximately 6.5 km/sec). This is in contrast to high densities ( $2.8\text{--}3.1 \text{ g/cm}^3$ ) and seismic velocities (estimated at approximately 6.8-7.2 km/sec) for the supracrustal belts. Overall, this crustal section is dominated by a low P-wave velocity matrix that surrounds discontinuous, high velocity layers, thus there is little seismic distinction between the upper, middle and lower crust. Work in progress (Shive and Williams) indicates that magnetite is also the magnetic phase in PGB enderbites but susceptibilities are much less than IVZ granulites, suggesting that lower crustal magnetic patterns will differ substantially for the two crustal cross-sections. Work on heat production is also in progress. Preliminary data indicate no significant difference in  $\text{K}_2\text{O}$  content between PGB enderbites and CLS granitic rocks suggesting there may be no major change in heat production across the amphibolite-granulite facies boundary.

The IVZ and SCZ represent Phanerozoic crust in which mafic and ultramafic magmas were added to deeper portions of a modestly metamorphosed pelitic package of sediments. Physical properties of mafic, ultramafic and de-granitized pelitic rocks are distinctively different than upper level granites and pelitic schists and gneisses resulting in a geophysically zoned crust. This zonation is enhanced by the Pogallo line, a Jurassic low-angle normal fault. In contrast, the granite-greenstone belt mode of evolution of the PGB and CLS created a crustal cross-section with little geophysical distinction between various crustal levels although significant lithologic heterogeneity exists. In the future, geophysical techniques may have the resolution necessary to detect differences such as these thus permitting routine analysis of crustal evolution through use of geophysical surveys.

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# USE OF OLIVINE AND PLAGIOCLASE SATURATION SURFACES FOR THE PETROGENETIC MODELING OF RECRYSTALLIZED BASIC PLUTONIC SYSTEMS

Hanson, Gilbert N., Earth and Space Sciences, SUNY, Stony Brook, NY 11794

During petrogenetic studies of basic plutonic rocks, there are at least three major questions to be considered: (1) what were the relative proportions of cumulate crystals and intercumulus melt in a given sample? (2) what is the composition and variation in composition of the melts within the pluton? and (3) what is the original composition of the liquids, their source and evolution prior to the time of emplacement? These questions are difficult to attempt to answer in unaltered bodies. They are even more difficult to answer in bodies which have undergone recrystallization and metamorphism as have most Archean basic plutons. In extreme cases one may not even be sure a unit represents a metamorphosed basic pluton. Use of the olivine and plagioclase saturation surfaces can help to answer some of these questions in recrystallized bodies.

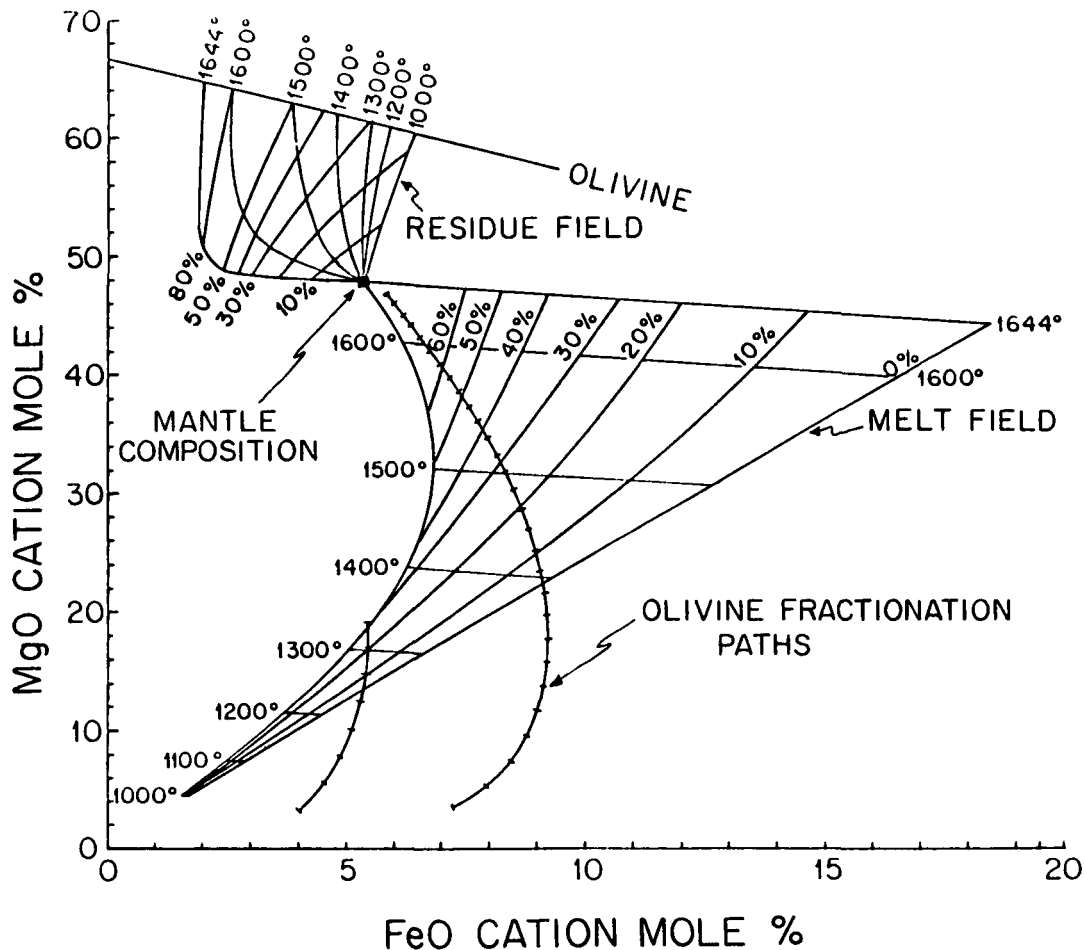


Figure 1.

# SATURATION SURFACES

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Use of the olivine saturation surface diagram of Roeder and Emslie (1) on which possible melts and residues of mantle melting can be plotted and contoured for extents of melting and temperature of melting (one atmosphere) for a given mantle composition (Fig. 1 from Hanson and Langmuir, 2) permits distinguishing whether a sequence of ultramafic rocks represents cumulates or residual mantle. A sequence of samples of residual mantle should plot close to the residue field in Fig. 1. The exact position of any sample will be dependent on the composition of the mantle, the extent and conditions of melting, and the fraction of melt retained with the residue.

A sequence of cumulates need not be restricted to the residue field in Fig. 1. If the composition of a cumulate plots close to the olivine composition, the range in compositions of the melt from which that cumulate precipitated must lie within the range of isopleths with the potential olivine compositions in the cumulate. For example in Fig. 2 for the cumulate plotted, the limits to the range of possible compositions for the cumulate olivine will be dependent on whether the rock represents an

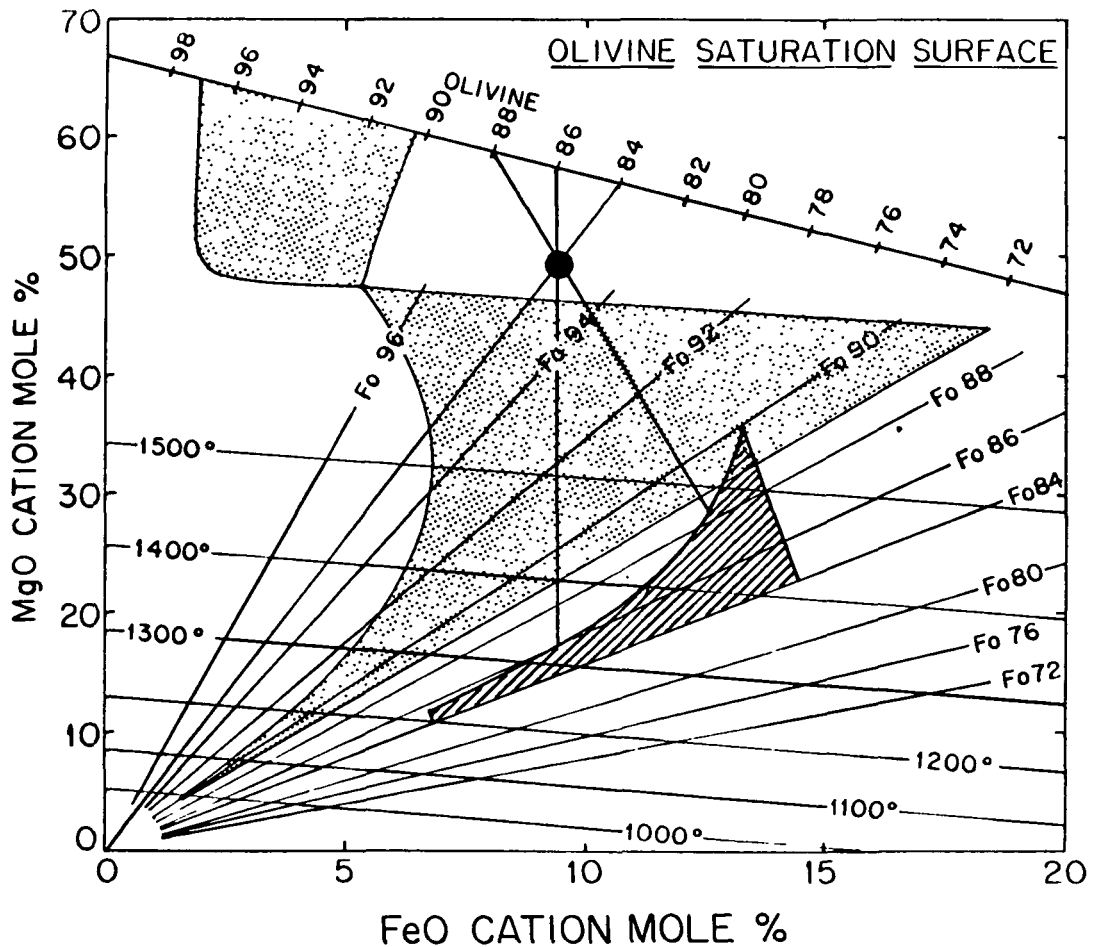


Figure 2.

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orthocumulate or an adcumulate. For a metamorphosed rock this may not be apparent. If it is an adcumulate, because the pyroxenes generally have similar or higher Mg numbers than olivine, the maximum forsterite content of the olivine is given by a line from the origin through the sample intersecting the olivine composition, in this example Fo-84. If the sample represents an orthocumulate, the composition of potential melts is given by each line from the compositions of olivine with Fo greater than 84, through the sample to the appropriate isopleth. Two lines are shown as examples in Fig. 2. The field of possible intercumulus melts shown are below the primary melt field as might be expected for melts which have undergone olivine fractionation and is similar to the fields for mafic Archean volcanics.

The plagioclase saturation surface is shown in Fig. 3 from Langmuir (3) in which cation normative An and Ab are used to represent the plagioclase components in basic to dacitic melts. The composition and temperature of crystallization of liquidus plagioclase are shown on their respective contours. The recovered composition and temperature are on average 3 mole % An and 12 degrees C respectively. Use of the olivine saturation surface in conjunction with the plagioclase saturation surface diagram allows consideration of whether a rock with a basaltic composition could have originally represented a melt. This is done by relating the composition of the rock to the possible melt field and olivine fractionation curves in Fig. 2 and comparing the liquidus temperatures for olivine and plagioclase on the respective saturation surfaces.

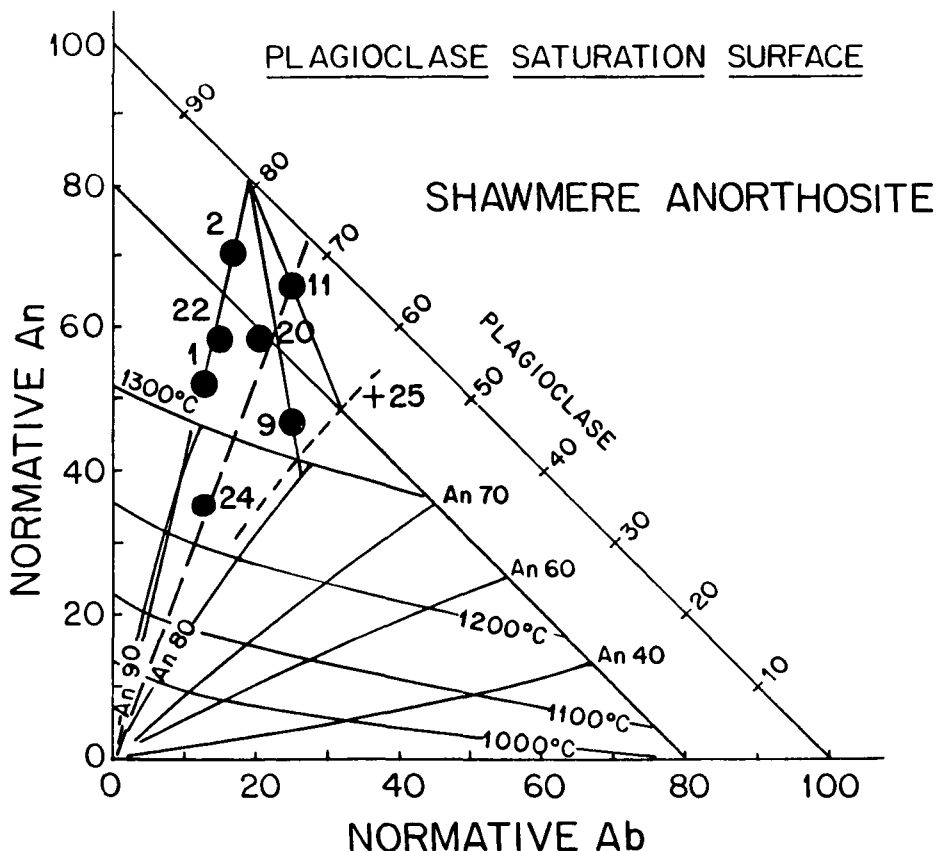


Figure 3.

The plagioclase saturation surface can also be used to place limits on the relative proportions and compositions of intercumulus melt and cumulate minerals in potential cumulate rocks (4). For example, cumulate samples from the Shawmere Anorthosite Complex are plotted in Fig. 3 as well as an anorthositic gabbro dike (Sample 25) from Simmons, Hanson and Lumbers (5). The cumulate minerals other than plagioclase plot near the origin of this diagram. Thus, any given rock must lie within the field defined by the origin, the cumulate plagioclase composition and the intercumulus melt composition. If samples 1, 2, and 22 are considered to be adcumulates, the cumulate plagioclase has a composition of An-81, which is the composition of recrystallized plagioclase from the core of a remnant megacryst in sample 2. If they contain significant intercumulus melt, the cumulate plagioclase has a higher An content. Thus the melt lies on an isopleth with an An content equal to or greater than An-81.

Assuming that sample 11 may have been derived from a melt similar to that from which sample 2 crystallized, we may use it to help place constraints on the normative plagioclase composition and abundance of the intercumulus melt. Sample 11 has a REE pattern and abundances compatible with significant cumulate plagioclase as does sample 2. Sample 11, however, has 2.3 to 4.0 times the abundance of REE for lights to heavies respectively compared to sample 2. This suggests that sample 11 is a mesocumulate or orthocumulate. Thus, the cumulate plagioclase is more anorthitic than An-73 the composition of cumulate plagioclase if sample 11 were an adcumulate. If for example, the cumulate plagioclase had an An content of An-81, the intercumulus melt must lie along the An-81 isopleth and to the right of a line drawn between the compositions of An-81 plagioclase and sample 11. Sample 25, the anorthositic gabbro dike, has a composition similar on this diagram to such a potential intercumulus melt. For a given composition melt and cumulate plagioclase it is possible to calculate the proportions of melt, plagioclase and total other cumulate phases using the lever rule.

Use of both saturation surfaces can place strong limits on the compositions of potential cumulate phases and intercumulus melts. Consideration of appropriate trace elements can indicate whether a sample is an orthocumulate, adcumulate or mesocumulate. Thus, when trace element and petrographic data are considered together with the saturation surfaces, it should be possible to begin to answer the three major questions given above, even for strongly recrystallized basic plutons.

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AN ANCIENT DEPLETED MANTLE SOURCE FOR ARCHEAN CRUST IN RAJASTHAN, INDIA: J.D. Macdougall, Scripps Inst. Ocean., La Jolla, CA 92093, K. Gopalan, Phys. Res. Lab., Ahmedabad, India, G.W. Lugmair, Scripps Inst. Ocean., La Jolla, CA 92093 and A.B. Roy, Univ. of Rajasthan, Udaipur, India

Precambrian rocks exposed in the state of Rajasthan in Western India comprise the Banded Gneiss Complex (BGC) and the Aravalli and Delhi supergroups which overlie the BGC. Based on systematic mapping by Heron (1) these divisions are believed to represent major orogenic cycles; in spite of some disagreements and revisions, Heron's work remains the basic framework for interpretation of these rocks. Heron (1) considered the BGC to be the oldest unit exposed in Rajasthan. Although this has been challenged by some later workers, there is a clear erosional unconformity between the BGC and overlying metasediments and metavolcanic rocks of the Aravalli sequence in several exposures along the western margin of the BGC. Virtually no modern chronologic data are available for the BGC. However, recent Rb-Sr work on the Untala Granite, believed to be intrusive into the BGC, gave a whole rock isochron age of  $2950 \pm 150$  my (Choudhary et al, ref. 2). We therefore undertook a geochemical and isotopic study of the BGC in the expectation that valuable data about a little-known segment of the earth's Archean crust would result.

We report here data from an initial set of BGC samples from the area east of the city of Udaipur. In this region the BGC comprises typical grey gneiss with variably abundant granitic and mafic components. We have concentrated our efforts to date on the mafic components which, based on chemical data, appear to be metavolcanic. All samples examined were recrystallized under amphibolite or upper amphibolite facies conditions. Pertinent chemical data for a small number of amphibolites analyzed so far are:  $\text{SiO}_2$ : 49-53%;  $\text{MgO}$ : 5.7-7.3%;  $\text{K}_2\text{O}$ : 0.24-0.50%; Ni: 106-140 ppm; Zr: 37-159 ppm. From Sm/Nd data, all amphibolites show small to moderate LREE enrichments.

A group of nine samples (3 gneisses, 6 amphibolites) define a Sm-Nd isochron giving an age of 3.5 AE with an initial ratio corresponding to  $\epsilon_{\text{JUV}}(T) = +3.5$ . Rb-Sr data for the same samples show a large amount of scatter and provide evidence for later metamorphic disturbance of the Rb-Sr system. Internal (mineral) isochrons for associated granites indicate that Sr reequilibration occurred in these rocks at 800-900 my (2,3). Structural evidence suggests at least three and possibly more major deformational episodes (e.g., Roy et al., ref. 4), so that evidence for this complex metamorphic history in the Rb-Sr system in the gneisses and amphibolites we have analyzed is not surprising.

The Sm-Nd data indicate that the amphibolites and grey gneisses are essentially cogenetic, and the 3.5 AE age must date the time of crust formation. The initial isotopic ratio for these rocks, corresponding to  $\epsilon_{\text{JUV}}(T) = +3.5$ , joins a growing number of examples of early crustal segments created with positive  $\epsilon_{\text{JUV}}(T)$  values. If the JUV (or CHUR) reservoirs truly represent the bulk earth, then the source regions for these rocks were depleted very early in earth history. This is perhaps not very surprising since large-scale melting and accompanying fractionation must have occurred in the outer parts of the early earth. However, an important point is that the depleted source must have been preserved over a substantial period of

time in order to allow differential evolution of  $^{143}\text{Nd}$  by as much as 3.5  $\epsilon$  units. The degree of fractionation required depends critically on the timing of fractionation. For example, for the Rajasthan data, the increase in Sm/Nd in the source compared to JUV would have to have been 16% if the differentiation occurred at 4.5 AE, and as high as 47% if it occurred at 3.8 AE. The corresponding  $\epsilon_{\text{JUV}}$  values in the same source today would range from +18 to +45, respectively. Clearly such high values are not observed in the major depleted reservoir being sampled today, the MORB source. Thus either the early depleted sources were localized residues or cumulates never again sampled, or, if genetically related to the present day MORB source, the rate of increase of  $^{143}\text{Nd}/^{144}\text{Nd}$  has slowed appreciably, possibly due to crustal recycling.

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# VENUS TOPOGRAPHY; CLUE TO HOT-LITHOSPHERE TECTONICS?

George E. McGill, Department of Geology and Geography, University of Massachusetts, Amherst, MA 01003

One of the fundamental problems of Archean Earth history is the style of tectonics, especially whether or not some form of plate tectonics occurred (1, 2). The feasibility of Archean plate tectonics revolves around the questions of lithosphere temperature, thickness, and buoyancy, and of crustal thickness and composition.

Venus has a very hot surface (470°C) and thus, by inference, a hot crust and lithosphere. All current models for the thermal structure of Venus predict a hot and thin lithosphere (3, 4), and most workers agree that this implies a positive buoyancy for the lithosphere (5, 4) and thus the absence of "trench pull", one of the favorite driving forces for plate tectonics on Earth. Only one small region, Ishtar Terra (Fig. 1) exhibits topographic features consistent with plate convergence; a high

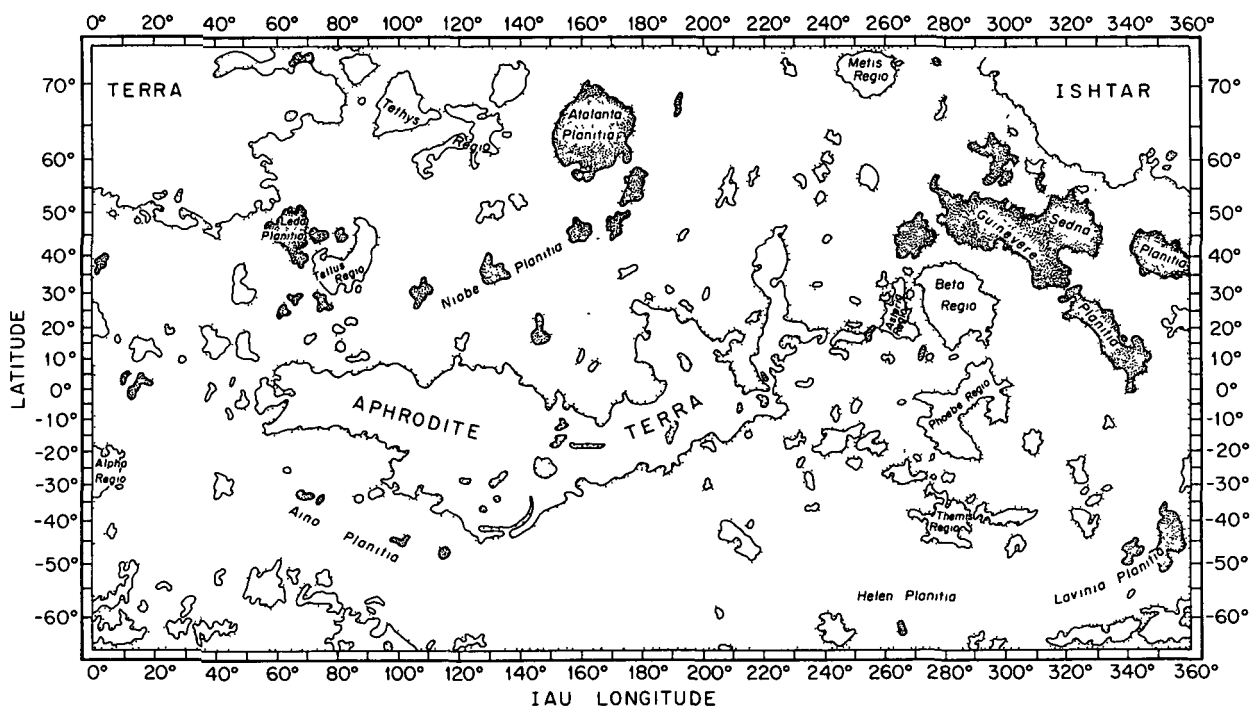


Fig. 1. Major physiographic regions of Venus. Heavy stipple, >1 km below median elevation; light stipple, median elevation  $\pm$ 1 km; unstippled, >1 km above median elevation.

plateau with marginal linear mountain belts (6). In contrast, there are extensive regions consistent with plate divergence (Beta Regio-Phoebe Regio, Aphrodite Terra, and others). These tend to be long, narrow, relatively elevated zones characterized by closed depressions with flanking elevations, relatively high



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rms slopes at meter scale, and relatively high surface roughness at centimeter scale (7). The closed-depression-flanking-high topography varies from long rift systems surprisingly similar in dimensions and general characteristics to continental rift systems on Earth (Fig. 2)(8), to shorter and less continuous rift-like features (Fig. 3), to irregular closed depressions and associated elevations of various sizes (Fig. 4). One

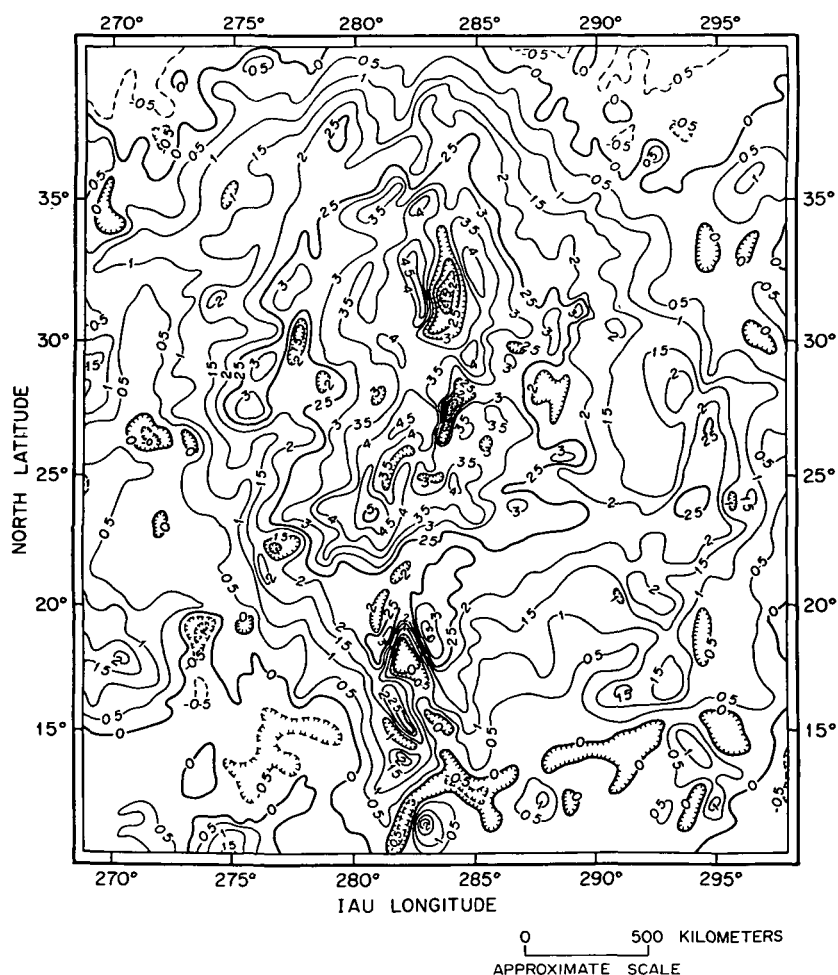


Fig. 2. Topography of Beta Regio  
Contour interval 0.5 km; datum is median elevation

possible tectonic style for a hot-lithosphere planet involves heat loss through numerous hot spots; regions of abnormally thin lithosphere, high conductive heat loss, and active volcanism (4). The abundance of closed depressions and associated elevations lying along linear elevated zones on Venus suggests that these "hot spots" are, in fact, concentrated along what amount to incipient divergent boundaries (9); regions of high heat flow, thermally elevated terrain, active volcanism, and limited (a few km) extension resulting in crustal collapse expressed as rifts or volcano-tectonic depressions. Except for the difference in  $H_2O$  abundances (Archean Earth was wet;

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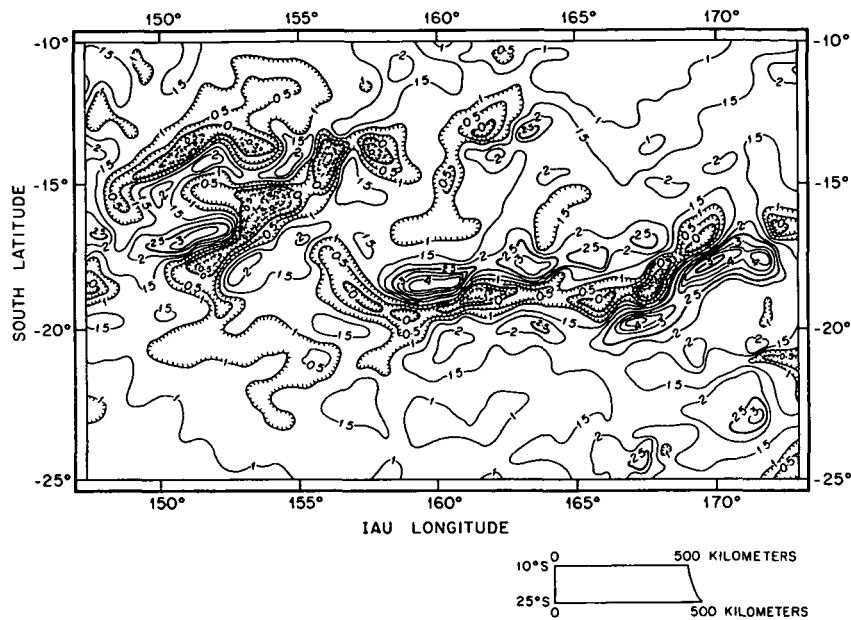


Fig. 3. Topography of central Aphrodite Terra. Contour interval 0.5 km; datum is median elevation.

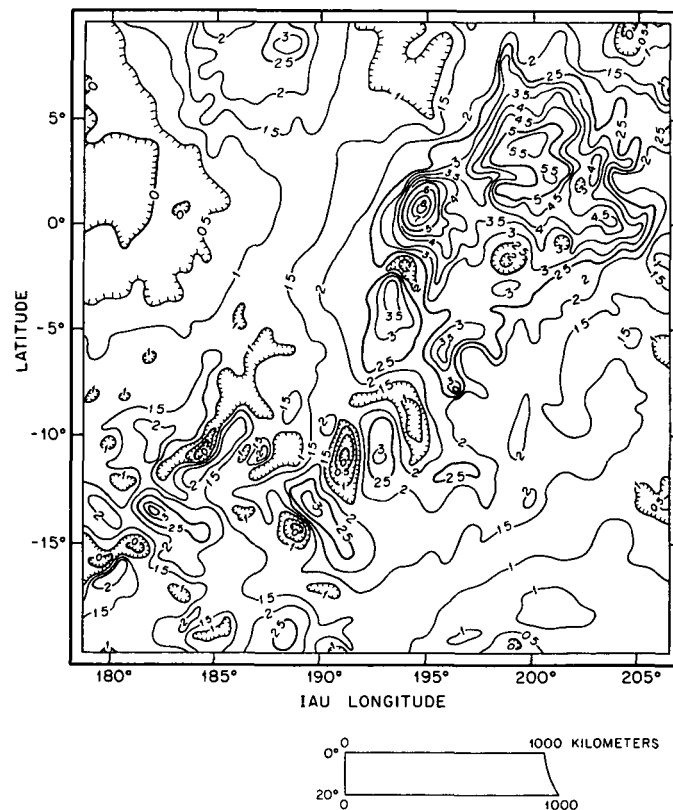


Fig. 4. Topography of eastern Aphrodite Terra. Contour interval 0.5 km; datum is median elevation.

modern Venus is very dry), these linear zones on Venus may be analogous to the tectonic settings for some Archean greenstone belts.

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# DEVELOPMENT OF THE EARTH'S EARLY CRUST: IMPLICATIONS FROM THE BEARTOOTH MOUNTAINS

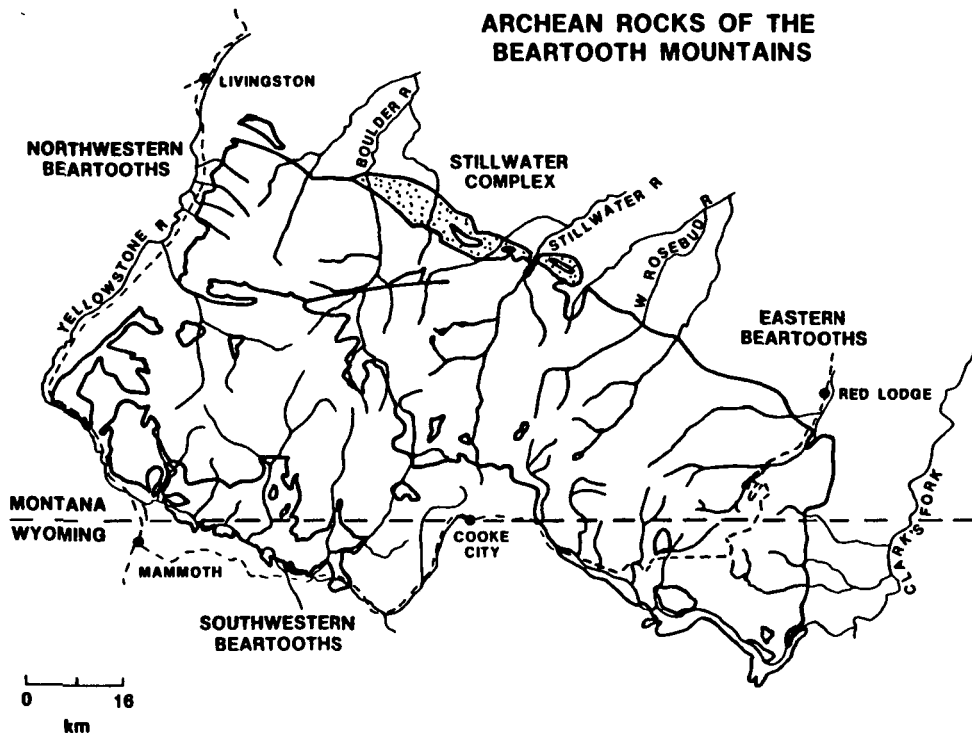
P. A. Mueller, Dept. of Geology, Univ. of Florida, Gainesville, FL 32611

J. L. Wooden, Lunar and Planetary Institute, Houston, TX 77058

D. J. Henry, Lunar and Planetary Institute, Houston, TX 77058

D. W. Mogk, Dept. of Earth Sci., Montana State Univ., Bozeman, MT 59717

**INTRODUCTION.** The Beartooth Mountains of Montana and Wyoming are one of several major uplifts of Precambrian rocks in the northwestern portion of the Wyoming Province. The range is composed of a wide variety of rock types which record a complex geologic history that extends from early(>3400 Ma) to late(~700 Ma) Precambrian time. The Archean geology of the range is complex and many areas remain unstudied in detail. In this discussion we will focus on two areas for which we have accumulated considerable structural, geochemical and petrologic information. The easternmost portion of the range(EBT) and the northwesternmost portion, the North Snowy Block(NSB), contain rather extensive records of both early and late Archean geologic activity. These data are used to constrain a petrologic-tectonic model for the development of continental crust in this area.

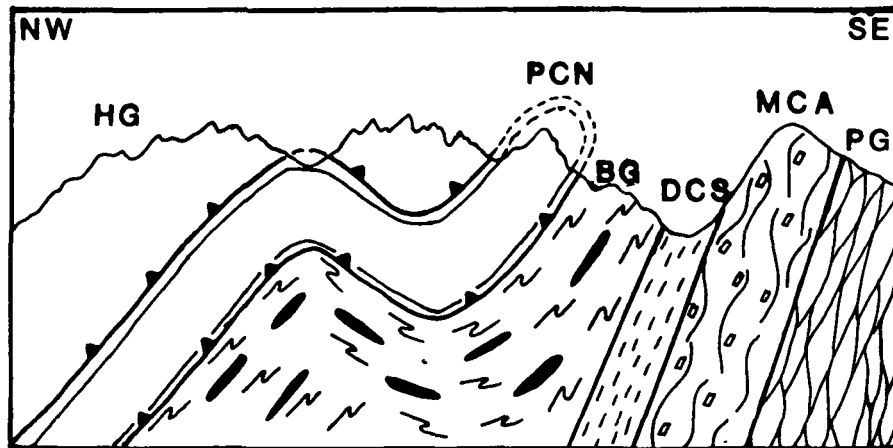


**GEOLOGIC FRAMEWORK. Early Archean** The oldest rocks identified in the EBT are supracrustal rocks that include metamorphosed basalts, silicic volcanics, ultramafics, ironstones, and various clastic sedimentary rocks(pelites, wackes, and quartzites)[1]. These rocks are found as meter- to several kilometer-sized inclusions in younger granitoids. The contacts between the different supracrustal rock types are usually tectonic and little original stratigraphy is discernible. Major, trace and REE analyses of these supracrustal rocks reveal several interesting features: 1) The earliest basaltic rocks differ from later ones in the EBT by having generally low REE abundances(10-20x) with relatively flat patterns; some komatiitic compositions may be present. 2) Rocks whose protoliths were intermediate to silicic volcanics are relatively abundant; those having dacitic to rhyodacitic compositions have REE patterns

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much flatter than later intrusive rocks of similar bulk compositions. 3) Clastic rocks typically show notably high Cr and Ni abundances, especially in the quartzites[2]. Larger enclaves of these supracrustal rocks record a granulite grade metamorphic event(800°C, 6 kb) while smaller enclaves and xenoliths are more likely to record a later amphibolite grade metamorphic event. Rb-Sr and Sm-Nd whole rock data for a variety of supracrustal rock types suggest that the time of granulite grade metamorphism was 3400 Ma ago[3].

Early Archean rocks from the NSB are concentrated in a single structural unit, the basement gneiss(BG of the cross section)[4]. This unit constitutes a ductile shear zone that formed at relatively low temperature(~500°C). Compositionally, the unit is typically trondhjemitic with amphibolitic layers conformable to the shear foliation. Rb-Sr and Sm-Nd studies indicate that the silicic portion of the gneiss is ~3500 Ma old. The relationship of the age of shearing to the age of formation of the protolith is unresolved. Other early Archean rocks are found in the heterogeneous gneiss(HG) unit of the cross section. This unit is a complex mixture of supracrustal lithologies and tonalitic to granitic gneisses. The tonalitic portions appear to be chronologically equivalent to the basement gneiss.



Cross section of the North Snowy Block. The lithologic units are the Heterogeneous Gneiss(HG), the Pine Creek Nappe(PCN), the Basement Gneiss(BG), the Davis Creek Schist(DCS), the Mount Cowan Augen Gneiss(MCA), and the Paragneiss(PG).

Late Archean The EBT was the locus of major magmatic and metamorphic activity during the period 3000-2800 Ma ago. This episode of intense activity began with the eruption and shallow level emplacement of substantial volumes of andesitic magmas, many of which are unusually enriched in incompatible elements[5]. Subsequently, these rocks were subjected to amphibolite grade metamorphic conditions about 2850 Ma ago[1]. During the latter stages of this metamorphic event small volumes of incompatible element-rich granodiorites and much larger volumes of more normal tonalitic to granitic melts were introduced[6]. Diabase dikes were intruded immediately after the emplacement of the granitoids and dike intrusion continued intermittently until ~700 Ma.

The NSB experienced a very different late Archean history. The main lithologic associations of the NSB, except for the MCA, are in tectonic contact with one another and these contacts mark discontinuities in metamorphic grade and structural style. It appears that four of these units(PG, BG, DCS and MCA) came into their present positions prior to the emplacement of the nappe units(PCN and HG). The PCN nappe complex contains an amphibolite grade assemblage of mafic amphibolite, schist, quartzite and marble. Limited Sm-Nd data

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suggest the amphibolite originally crystallized ~3000 Ma ago. Magmatic activity is represented by the Mt Cowan gneiss(MCA) which is now a K-spar augen gneiss that was apparently intruded close to its present position. Other than the MCA, none of the other lithologic packages can be demonstrated to be autochthonous.

**GEOLOGIC-TECTONIC MODEL.** Any reconstruction of the sequence of events that led to the development of the current ~45 km thick continental crust in this portion of the Wyoming Province must rely heavily on data from the Beartooth Mountains. Geochemical and geochronologic information for terrains to the west is not abundant; particularly with regard to the nature and distribution of early Archean rocks. The following model, therefore, accommodates the data from the Beartooths and is compatible with the limited information available for the surrounding terrains. A proposed scenario is depicted in a sequence of cartoons at the end of this section.

The Early Archean At 3400 Ma the EBT was already a well developed continental terrain with ultramafic to silicic components that were the source for the wide variety of supracrustal rock types subjected to granulite grade metamorphism. The environment of deposition was most likely a subsiding shelf bordering a continental mass (Beartoothia?) of unknown size and thickness. That these shelf sediments were subjected to granulite grade conditions and experienced recumbent isoclinal folding suggests tectonic burial of Beartoothia by collision with another continental mass. It appears that this other continent lay to the west and was at least 20 km thick. Assuming that continental masses in this area were of comparable thickness, it seems likely that pre-collision Beartoothia was at least 20 km thick and that by 3400 Ma was roughly 40 km thick with a metamorphic geotherm comparable to modern orogenic areas (i.e. ~40°C/km). Because of the lack of variety in early Archean rocks in the NSB, they offer little additional insight into the picture developed from the EBT data except to demonstrate the relatively widespread nature of the ~3400 Ma crustal component.

Late Archean The more abundant late Archean rocks in both the EBT and NSB clearly point to differences in the tectonic regimes of the two areas. The EBT was relatively quiet from 3400 Ma until roughly 3000 Ma when large volumes of intermediate magmas were generated[6]. This occurrence is distinctive because large volumes of rocks of intermediate composition are rarely found outside of greenstone belts. Their presence here is ascribed to subduction tectonics associated with closure of an ocean basin that lay to the west. The rocks were then subjected to amphibolite grade conditions ~2850 Ma ago[1] and subsequently intruded by a suite of late synkinematic to post-kinematic granitoids(2800 Ma)[6]. Pressure estimates based upon the compositions of diabase dikes that were intruded immediately after the emplacement of the granitoids suggest crystallization at ~5 kb[8]. This magmatic episode apparently represents the end point of a major crust forming cycle as essentially no further diastrophic activity occurred in this area until Laramide time when the block was uplifted.

The late Archean history of the NSB also appears to have begun 3000 Ma ago with the extrusion of the mafic rocks of the PCN unit. Other members of this unit(quartzites, schists and marbles) suggest that the sequence was laid down on a continental margin or extensional basin underlain by continental crust. The metasedimentary nature of rocks to the west of the NSB suggests a continental margin may be more likely[10]. Compression of this area was probably concomitant with the magmatic and metamorphic activity in the EBT and terminated with the emplacement of the major thrust sheets(PCN+HGN). The allochthonous nature of many of the major lithologic packages in the NSB is reminiscent of modern accreted terrains and it is possible that major strike-slip faults were generated along this continental margin during late Archean time. Such fault systems could have moved crustal masses from the west into their

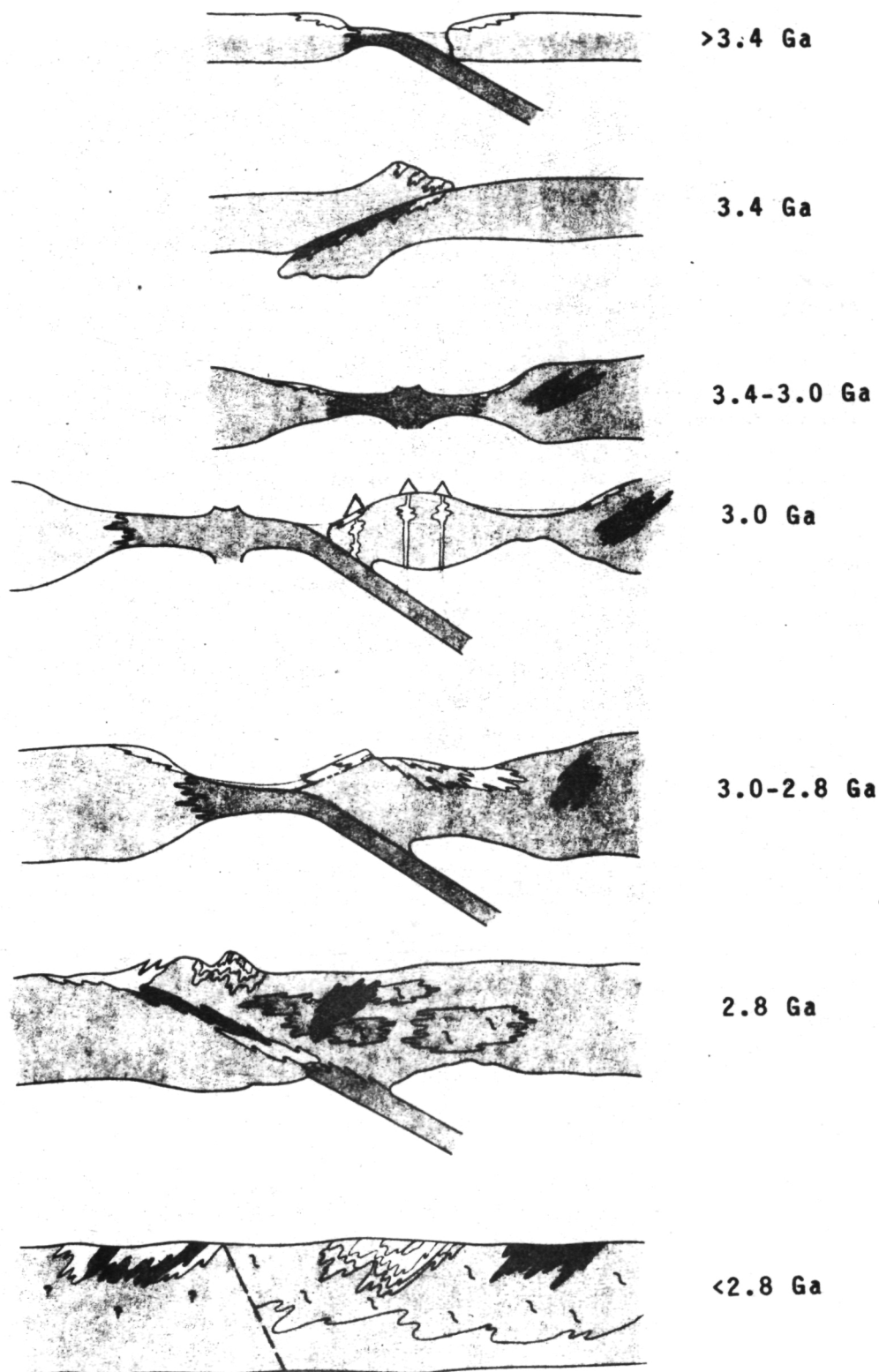
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present positions. The NSB, therefore, may mark the eastern limit of an Archean analog of a modern Cordilleran margin.

**IMPLICATIONS.** It is clear from the foregoing discussion that two very different types of late Archean terrains are present in the NSB and EBT regions of the Beartooth Mountains. Moreover, a review of the regional Archean geology shows that the Beartooth Mountains may lie astride a fundamental crustal boundary in the Wyoming Province. The NSB and terrains generally to the north and west are composed primarily of metasedimentary rocks with major thrust and strike-slip faulting[9,10,11]. The EBT and terrains generally to the south and east are dominated by late Archean magmatic rocks[9]. Our present interpretation suggests that this boundary was an Archean continental margin similar to modern complex continental margins that have experienced the accretion of allochthonous terrains and calc-alkaline magmatism in close proximity (e.g. the late Phanerozoic of the northwestern United States). If this interpretation is correct, the accreted nature of the Archean continental base of North America west of the Beartooths is an unevaluated aspect of the complex late Phanerozoic geologic history of the region.

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**Tectonic Model for Archean Crustal Evolution in SW Montana**





CLASTIC SEDIMENTARY ROCKS OF THE MICHIPICOTEN VOLCANIC-  
SEDIMENTARY BELT, WAWA, ONTARIO. Richard W. Ojakangas, Department of  
Geology, University of Minnesota, Duluth, MN 55812

The Wawa area, part of the Michipicoten greenstone belt, contains rock assemblages representative of volcanic-sedimentary accumulations elsewhere on the shield. Three mafic to felsic metavolcanic sequences and cogenetic granitic rocks range in age from  $2749 \pm 2\text{Ma}$  to  $2696 \pm 2\text{Ma}$ <sup>1</sup>. Metasedimentary rocks occur between the metavolcanic sequences. The total thickness of the supracrustal rocks may be 10,000 m. Most rocks have been metamorphosed under greenschist conditions. The belt has been studied earlier<sup>2-5</sup> and is currently being remapped by Sage<sup>6</sup>.

The sedimentologic work has been briefly summarized<sup>7</sup>; two main facies associations of clastic sedimentary rocks are present - a Resedimented (Turbidite) Facies Association and a Non-marine (Alluvial Fan-Fluvial) Facies Association.

The Resedimented Facies Association consists of conglomerates, graywackes and mudstones. The sedimentary sequence is thick in the west and passes into thin carbonaceous and pyritic shales to the east; iron-formation forms a marker horizon in the sequence, passing from iron-oxides in the west where the conglomerates are thick, to carbonate in the vicinity of the felsic volcanic centers, to sulfide facies to the east in the shale<sup>4</sup>. The graywackes have the classic characteristics of turbidites, including excellent grading and internal Bouma intervals; the interbedded mudstones, with minor silty laminae, indicate deposition under low energy conditions (Fig. 1). The conglomerates associated with the graywackes in the west have an abundant poorly-sorted matrix of sandy graywacke or chlorite-shale, and are crudely stratified<sup>4</sup>. The assemblage is indicative of deposition on submarine fans, although slumping from the edges of explosive volcanic edifices does not necessitate the presence of fans.

The Non-Marine Facies Association consists largely of cross-bedded lithic to feldspathic sandstones with local conglomeratic units. Sedimentary features include small- to medium-scale cross-bedding of both trough and planar types (Fig. 2), parallel bedding, parting lineation, ripple marks, soft sediment deformation (water escape structures?), mudcracks, and mudchip horizons. The assemblage is characteristic of braided fluvial and alluvial fan sequences (and perhaps deltaic), rather than meandering river channel-floodplain sequences which would contain more muddy and silty units.

Mapping by Sage<sup>6</sup> has revealed that locally the sequence consists, from the bottom up, of volcanics, turbidites, cross-bedded sandstones, and conglomerates. Sage (personal communication, 1981) has suggested this constitutes a shoaling-upward sequence as the depocenter was locally filled.

Preliminary petrographic studies indicate that both facies associations consist largely of felsic volcanic debris (Table 1). Recrystallization has commonly partially obscured original textures, but in the least metamorphosed samples detrital grains are readily discernable (Figs. 3 & 4). Clearly, felsic volcanic sources were dominant; in addition to the volcanic rock fragments, most of the fine-grained, recrystallized quartz-feldspar matrix and at least part of the quartz and plagioclase may also have been derived from volcanic sources. Plutonic

detritus is minor and may reflect derivation from coeval plutons as suggested earlier<sup>4</sup>. The Q:F:L (quartz to feldspar to lithics or rock fragments) ratios for the point - counted samples of Table 1 are 23:23:54, 13:15:72, and 31:12:57. If the fine quartz-feldspar matrix is assumed to have originally been sand-sized rock fragments, the L component increases to 58, 78 and 64. Much of the sediment may have been derived from the reworking of unconsolidated or poorly consolidated pyroclastic detritus, as suggested by Ayres<sup>8</sup> for graywackes of the Gamitagama belt 50 km to the south and by Ojakangas<sup>7</sup> for much of the sedimentary detritus in Archean volcanic-sedimentary belts of the Canadian Shield.

Additional work is planned in the Wawa area to better determine the paleogeography, including the temporal and spatial relationships of the turbidites and other sandstones to each other and to the volcanic rock units. Pyroclastic rocks are abundant, and probably were deposited in both subaerial and subaqueous environments<sup>9,10</sup>. Transitions should exist between the pyroclastic and the sedimentary rock units. Because of structural complications, including overturned folds and differential movement of fault blocks, detailed sedimentological studies have of necessity awaited field mapping in progress by Sage.

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TABLE 1

MODAL ANALYSES

	Turbidite facies ("Graywacke") WA-81-8B	Trubidite facies ("Graywacke") WA-81-10B	Fluvial facies ("Sandstone") WA-81-11
Rock Fragments:			
Volcanic, felsic	33	39	31
Volcanic, intermediate-mafic	4	8	1
Plutonic, feldspar-quartz	4	3	5
Quartz			
Common	4	4	13
Polycrystalline, recrystallized	14	5	7
Feldspar	18	10	8
Matrix-cement			
Micas	8	13	18
Quartz-feldspar (fine)	10	17	13
Epidote	2	1	1
Carbonate	4	-	3



Figure 1. Interbedded graywackes and mudstones near the dam on the Magpie River, Chabanel Township, north of Wawa.



Figure 2. Cross-bedded sandstones near Bauldry Lake in Esquega Township northeast of Wawa.

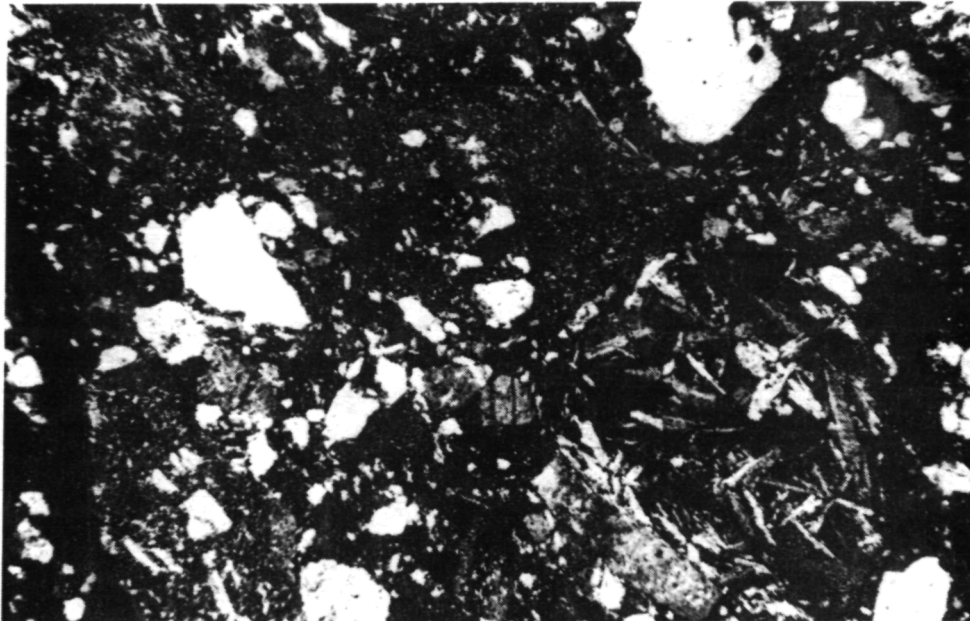


Figure 3. Photomicrograph of representative graywacke. Note mafic volcanic fragment at lower right, felsic volcanic fragments at lower left and upper center, plagioclase, and common quartz grains. Field of view is about 2.5 mm high. Crossed nicols.

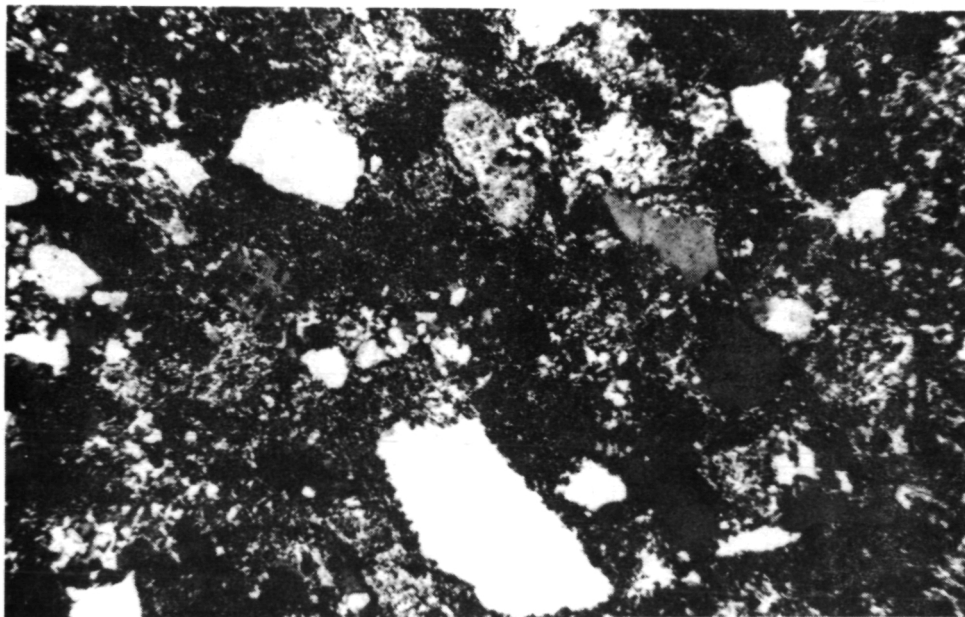


Figure 4. Photomicrograph of representative cross-bedded sandstone. Most grains are felsic volcanic fragments of various textures and common quartz. Field of view is about 1.5 mm high. Crossed nicols.

# A MULTI-ELEMENT STUDY OF ISUA IRON-FORMATION, W-GREENLAND.

Ullrich Rast, Max-Planck-Institut für Chemie, Abt. Kosmochemie,  
Saarstrasse 23, D-6500 Mainz, F.R. Germany.

Meta-sediments from Isua, West-Greenland were analyzed by instrumental thermal neutron activation analysis (ITNAA). These sediments are chemical precipitates having some layers of remarkably high Cr content. The latter were compared to Cr-poor layers. It turned out that the Cr-enriched layers had higher Ir- and Ni-contents than the samples from the Cr-poor layers. Compared to phanerozoic samples the highest Ir-contents are not extraordinarily higher than in a "modern" sediment, and the Cr-poor layers, representing more or less "normal" sedimentation are not significantly higher in Ir than an average phanerozoic shale (s. table).

From the cratering record of the moon we can assume a similar cratering of the Earth at about the time when the Isua rocks were formed. Can such an enormous flux of extraterrestrial material be detected in those very old Archean rocks? The noble metals, esp. Ir, prove to be sensitive tracers for the detection of an extraterrestrial component in the Earth's crust because of their high depletion in the crust relative to chondritic abundances. Consequently analyzing the noble metal content of sedimentary rocks of archaic, proto- and phanerozoic age one should find a certain decrease in noble metal content from the archaic to "modern" sediments, representing the decrease of the total amount of the "late accretion" component. When looking at the data (s. table) from the Isua rocks one cannot find that the amount of Ir is considerably higher than in an average phanerozoic sediment taking into account that the Cr-poor layers don't show an Ir enrichment relative to "modern" sediments at all! The question is now are the Isua metasediments still too young to find this higher proportion of a meteoritic component which would mean that the time of the heavy meteorite bombardment ended before the Isua sediments were deposited. Or is there any geochemical or geodynamic process responsible for a redistribution, mixture or dispersion of an extraterrestrial component added to the Earth's crust, thus making its detection difficult or impossible? From the geochemical properties of many noble metals we can assume that they are not very mobile under sedimentary and low grade metamorphic conditions. If a greater proportion of the noble metals in the crust are brought to crust from the space they should be found in a roughly chondritic ratio and be clearly detectable from the highly fractionated indigenous siderophiles (cp. HERTOGEN et al., 1980). Using the Ni/Ir ratio for tracing meteoritic material, we must consider the mobility of Ni in low temperature geological processes, a difficulty which became evident in the discussion of the C/T-boundary noble metal enrichments. As shown by PALME (1982) impact melts of terrestrial meteorite craters contain several noble metals nearly unfractionated relative to chondritic abundances. But not all impact-generated materials have such a quality of preservation as the 290 m.y. old Clearwater impact melt. The "boundary clays" of the C/T-boundary which are thought to contain a certain proportion of meteoritic material are much younger but the original pattern is far less well preserved than the impact melt (PALME, 1982). And all these materials are much younger than archaic rocks representing sudden and drastic events while in the archaic rock, we are looking for an evidence of a "background" continuous flux. Nevertheless it is clear that late accretional processes have brought a small amount of Ir to the Earth's upper mantle, at a time when this amount of siderophile elements could not be extracted by a metal phase and brought to the core. At this time the mantle should have been roughly chondritic in the noble metals. Partial

melting, differentiation and crust formation created rocks which were highly depleted in Ni and Ir bearing phases. Despite the extrusion of considerable quantities of unfractionated mantle material, may be partly after massive break-up of the early crust as a result of several giant meteorite impacts, it is rather likely that the continuous flux of small extraterrestrial objects and dust were the main source of most of the noble metals contained in the archaean crust. But a lot of geological processes superimposed a terrestrial signature on the abundance patterns which makes it so difficult to account for the meteoritic component in the old rocks.

The Cr-rich layers of the Isua rocks contain not only a siderophile-enriched component but they are also enriched in Sc, Hf, Ta, and to a much lesser extent in other lithophiles like Th, U and the REE esp. the light REE. The element abundances of these samples are complex and point to different source materials. APPEL (1979) proposed a meteoritic origin of chromite grains found in these layers but the chemistry of the host rock does not show any features compatible with this idea.

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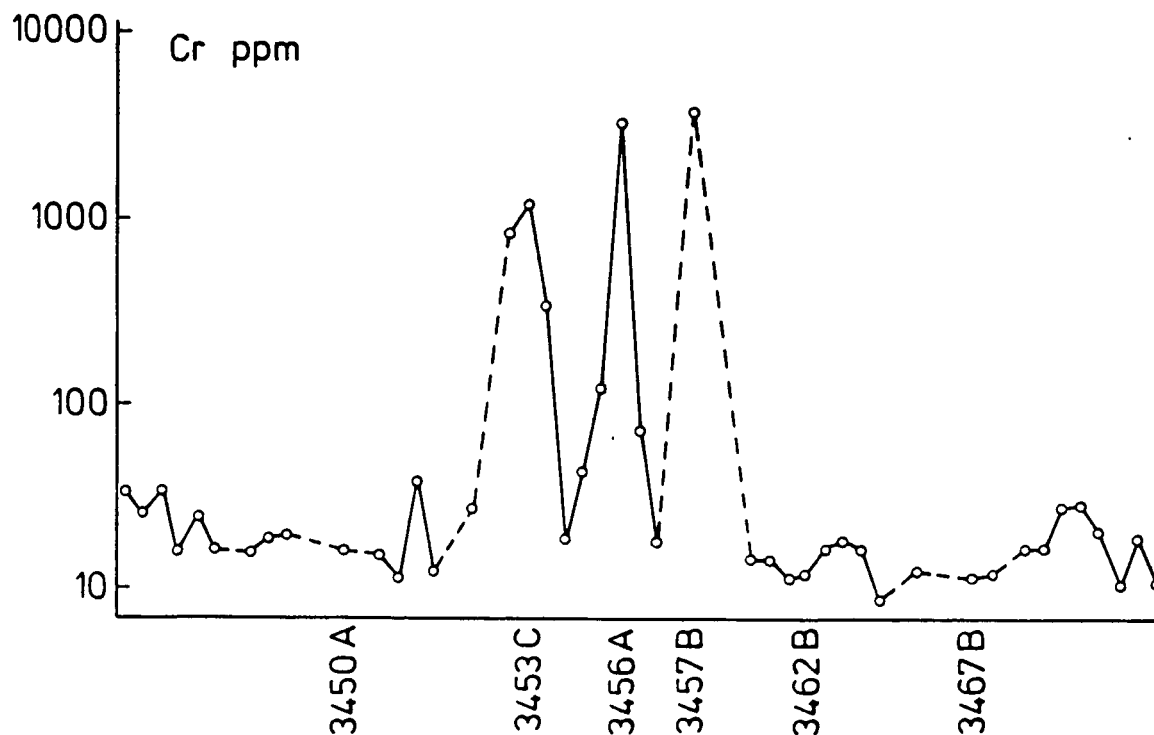


Fig. 1: Chromium-content of a section in the Isua iron-formation. Unpublished data obtained by atomic absorption spectroscopy, kindly permitted for use by P. APPEL, Geological Institute, University of Copenhagen. Samples analyzed by ITNAA are marked by sample number.



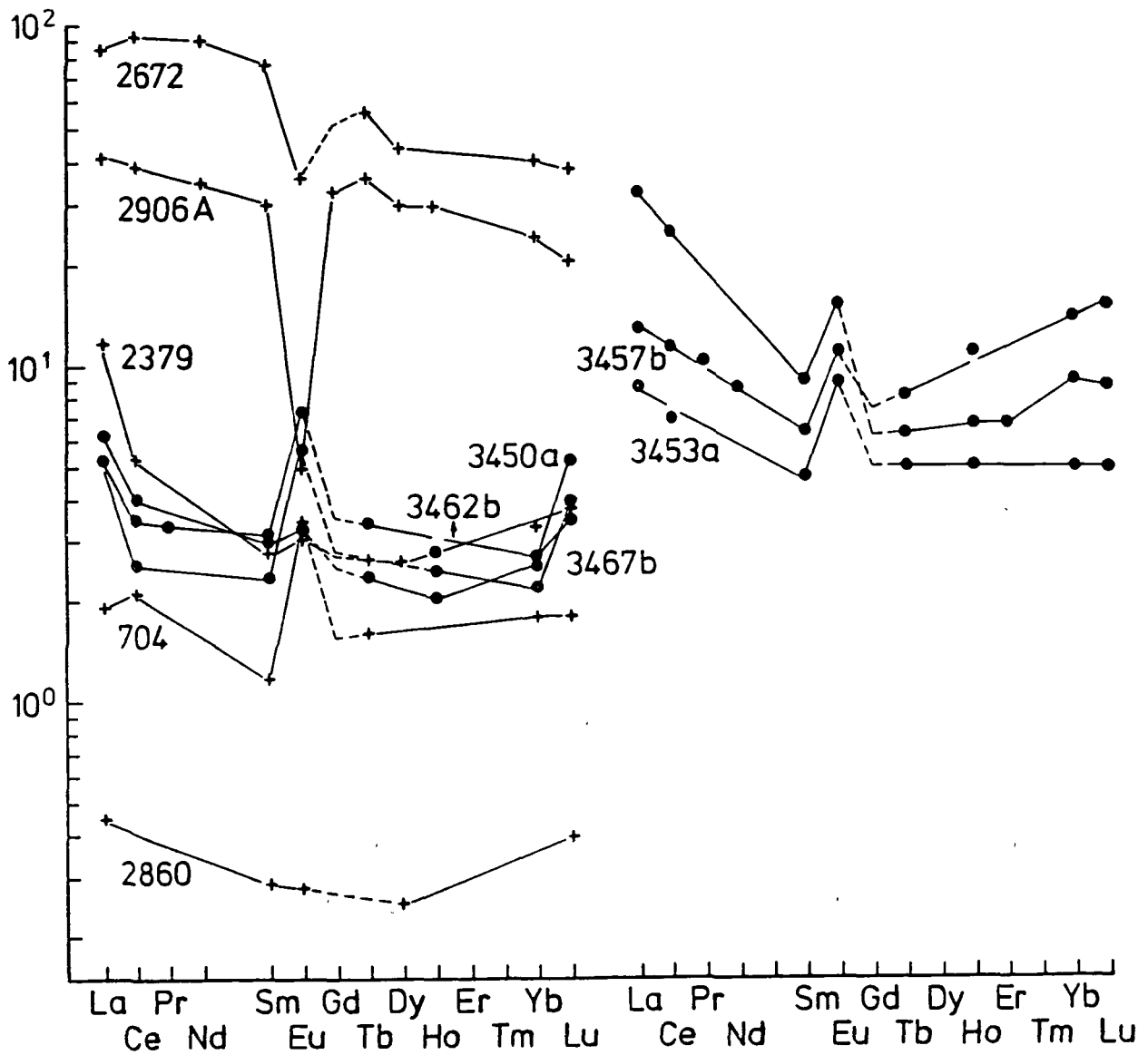


Fig 2: REE-pattern of Isua-rocks normalized to C1-type carbonaceous chondrite REE contents. Data from samples 2860, 704, 2379, 2906A, 2672 unpublished results from B. SPETTEL, Max-Planck-Institut f. Chemie, Mainz, kindly permitted for use. The REE values of C1-type carbonaceous chondrite were taken from PALME et al. (1981).

Rast, U.

Sample	Na	K	Rb	Cs	Sc	Cr	Mn	Fe[%]	Co	Ni	Cu	Zn	Ga	As	Br	Sb
3450a	16.8	6	36	<0.05	2.68	4.91*	31400	46.4	17.7	72*	44.5	63.3	3*	0.09	0.3	0.03
3453c	27.4	17		<0.1	24.8	1410	22800	36.8	34.5	400	38	230	14.6	0.345	0.2	0.14
3456a	29.2	12			24.2	3590	18900	42.3	27.3	735		153	15.5	0.345	0.2	0.13
3457b	28.0	10	26	<0.14	24.8	3940	9840	42.5	29.2	850		210	14.0	0.25	0.2	0.06
3462b	13.2	3	25	<0.24	2.55	8.24*	28200	48.2	20.4	77*	37.4	65.5	1*	0.048	0.2	0.06
3467b	15.8	5	50	0.2	2.68	9.59*	32100	44.5	16.2	70*	19.2	64.8	3.5*	0.165	0.2	0.04
Error [%]	3	30	30	50	2	2 <sup>2</sup> *7	2	2	2	2 <sup>2</sup> *15	15	10	5 <sup>5</sup> *15	25	25	30

Sample	La	Ce	Pr	Nd	Sm	Eu	Tb	Ho	Tm	Yb	Lu	Hf	Ta	Ir	Au	Th	U
3450a	1.53	2.58			0.46	0.19	0.087	0.116	0.14	0.42	0.13		0.01*	0.001	0.001	0.18*	0.14*
3453c	2.13	4.41			0.713	0.514	0.183	0.282	0.151	0.81	0.121	0.502	0.102	0.004		0.28	0.21
3456a	7.82	15.7		4.7	1.38	0.877	0.228	0.38		1.46	0.214	3.63	0.363	0.005	0.003	1.57	0.756
3457b	3.13	7.2		3.6	0.971	0.632	0.3	0.62	0.46	2.24	0.371	4.03	0.458	0.0033	0.004	1.89	1.09
3462b	1.28	2.25	0.32		0.485	0.433	0.127	0.16		0.45	0.088	0.03	0.18*	<0.001	0.002	0.065*	0.098*
3467b	1.30	1.65			0.361	0.308	0.23	0.14		0.36	0.098	<0.06		<0.001	0.0017	0.05*	0.15*
Error [%]	3	20	25	25	5	5	15	20	25	15	15	5	10 <sup>10</sup> *25	25	25	10 <sup>10</sup> *25	10 <sup>10</sup> *25

Tab. 1: Analytical results of several samples from the Isua iron-formation (cp. Fig. 1) obtained by ITNAA. The analytical precision is given in %. All values in ppm, except Fe [%].

# METAMORPHIC FLUIDS AND UPLIFT-EROSION HISTORY OF A PORTION OF THE KAPUSKASING STRUCTURAL ZONE, ONTARIO, AS DEDUCED FROM FLUID INCLUSIONS

R.L. RUDNICK\*, L.D. ASHWAL, and D.J. HENRY, Lunar and Planetary Inst. , Houston, TX 77058 (\*now at Res. School of Earth Sci., Aust. Nat. Univ., Canberra, Aust.)

## Introduction

Fluid inclusions can be used to determine the compositional evolution of fluids present in high grade metamorphic rocks (Touret, 1979) along with the general P-T path followed by the rocks during uplift and erosion (Hollister et al., 1979). In this context, samples of high-grade gneisses from the Kapuskasing structural zone (KSZ, Fig. 1) of eastern Ontario were studied in an attempt to define the composition of syn- and post-metamorphic fluids and help constrain the uplift and erosion history of the KSZ. Recent work by Percival (1980), Percival and Card (1983) and Percival and Krogh (1983) shows that the KSZ represents lower crustal granulites that form the lower portion of an oblique cross-section through the Archean crust, which was up-faulted along a northeast-striking thrust fault. The present fluid inclusion study places constraints upon the P-T path which the KSZ followed during uplift and erosion.

## Occurrence, Morphology and Composition of Fluid Inclusions

Fluid inclusions present in quartz in high-grade (700-800°C, 6-8 kbar) rocks (paragneisses, amphibolite, gabbro gneiss and a tonalite dike) collected near the Shawmere anorthosite complex in the KSZ (Fig. 1), consist of three types (listed in order of decreasing abundance): (1) CO<sub>2</sub>-rich inclusions (no visible H<sub>2</sub>O) (generally 1 to 12 µm, but up to 20 µm); (2) H<sub>2</sub>O-rich inclusions (no visible CO<sub>2</sub>) (1-35 µm); and (3) mixed CO<sub>2</sub> and H<sub>2</sub>O (of variable sizes).

CO<sub>2</sub>-rich inclusions occur along healed fractures and exhibit irregular to negative crystal morphologies with a few possessing an acicular morphology (up to 30 µm by 2 µm) (Fig. 2). At room temperature some of the CO<sub>2</sub> inclusions contain a birefringent solid phase that exhibits a large variation in relief upon rotation of the microscope stage (probably carbonate). In addition, there are acicular carbonate grains associated with acicular CO<sub>2</sub> inclusions within the same fracture. These CO<sub>2</sub> inclusions have apparently developed in the casts of carbonate grains. Melting points of the CO<sub>2</sub> inclusions range from -61.5 to -56.6°C (Fig. 3), indicating the presence of variable amounts of another component which depresses the melting temperature. Laser Raman spectroscopy performed on a CO<sub>2</sub>-rich inclusion which possesses one of the lowest melting temperatures (-61.5°C), shows the presence of CH<sub>4</sub> and no apparent N<sub>2</sub>. From this data, the melting point depressions of the CO<sub>2</sub> are tentatively attributed to varying amounts of CH<sub>4</sub> in the CO<sub>2</sub> phase. The amount of CH<sub>4</sub> within CO<sub>2</sub>-rich inclusions appears to be dependent on host rock lithology: meta-igneous rocks contain predominantly pure CO<sub>2</sub> (T<sub>m</sub> = -56.8 to -57.0 ± .5°C, with one trail in SH80-22A yielding a T<sub>m</sub> of -57.7°C, Fig. 3) while metasedimentary rocks contain varying proportions of CH<sub>4</sub> (T<sub>m</sub> = -61.5 to -57.2 ± .5°C, Fig. 3). Homogenization temperatures for the CO<sub>2</sub> inclusions (T<sub>h</sub>, vapor to liquid) range from -47 to +31°C (Fig. 3), with older-looking inclusions having lower T<sub>h</sub> than younger-looking inclusions. A late-stage tonalite dike (41-D2) and a garnet gabbro gneiss (22A) contain "pseudo-secondary", negative crystal form CO<sub>2</sub> inclusions (cf. Roedder, 1980), which are believed to have been entrapped during initial crystallization of the host mineral. The pseudo-secondary inclusions in the tonalite dike along with planar, negative crystal form inclusions in an amphibolite (42F), exhibit the lowest T<sub>h</sub> (highest density) of any CO<sub>2</sub> inclusions found in the KSZ rocks (Fig. 3). The high density inclusions in the amphibolite show significant melting point depressions, indicating the presence of CH<sub>4</sub>, which will cause T<sub>h</sub> to be lower than if the inclusion were pure CO<sub>2</sub>. Consequently, the densities of these inclusions are not as high as they appear, and the corresponding isochores are not representative of the actual P and T of entrapment. The CO<sub>2</sub> inclusions in the tonalite dike are relatively pure CO<sub>2</sub> (as seen by their melting temperatures), therefore, the T<sub>h</sub> yields an accurate density and the corresponding isochore can be used to determine the P or T of entrapment. The isochores for these high density inclusions in the tonalite pass through the lower portion of the T and P conditions of metamorphism estimated by Percival (1980, in press) (Fig. 4). This, plus the pseudo-secondary nature of the inclusions in the tonalite suggest that the CO<sub>2</sub> was trapped as the quartz crystallized during the granulite facies metamorphism.

H<sub>2</sub>O-rich fluid inclusions have been found in all lithologies studied. These aqueous inclusions are always in planar arrangements and have morphologies varying from irregular to ovoid to partial negative crystal form (Fig. 2). The planes of aqueous inclusions often cut across grain boundaries, indicating post-crystallization entrapment. The aqueous inclusions generally possess one or more daughter phases: several cubic, isotropic phases (NaCl plus ?), a

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rectangular, birefringent phase identified as  $\text{CaCO}_3$  through Raman spectroscopy and, rarely, an opaque, acicular phase. In addition, many  $\text{H}_2\text{O}$ -rich inclusions contain minute amounts of  $\text{CO}_2$ , which can only be observed through the formation of a clathrate which melts around  $+10^\circ\text{C}$ . Melting points for the  $\text{H}_2\text{O}$ -rich inclusions range from  $-37$  to  $+10^\circ\text{C}$  depending upon the dissolved components present, while some of the aqueous inclusions do not appear to freeze down to  $-190^\circ\text{C}$ .

Mixed  $\text{CO}_2 + \text{H}_2\text{O}$  inclusions (with both phases visible) are rare in the KSZ rocks and generally only occur where a trail of  $\text{H}_2\text{O}$ -rich inclusions intersects a trail of  $\text{CO}_2$ -rich inclusions. The morphologies of the mixed  $\text{CO}_2 - \text{H}_2\text{O}$  inclusions vary from negative crystal form (inherited from the original  $\text{CO}_2$  inclusions) to irregular.

#### Source of Fluids

The source of  $\text{CO}_2$  in granulite facies rocks is poorly constrained. Two models are generally invoked: (1)  $\text{CO}_2$  is derived from surrounding rocks by decarbonation reactions during metamorphism, or by oxidation of graphite, or (2)  $\text{CO}_2$  is derived from the mantle. Either or both of these two models may apply to the  $\text{CO}_2$  inclusions in the KSZ. The presence of  $\text{CO}_2$  filling carbonate mineral casts suggest that some  $\text{CO}_2$  may be derived from in situ decarbonation. However, the lack of extensive carbonate layers in the KSZ requires an additional source for the  $\text{CO}_2$ ; either unexposed carbonate layers, oxidized graphite (graphite occurs in some of the KSZ paragneisses (Percival, 1980)), or perhaps the  $\text{CO}_2$  is fluxed from the mantle (Newton et al., 1980).

The  $\text{H}_2\text{O}$ -rich inclusions and mixed  $\text{H}_2\text{O}-\text{CO}_2$  inclusions clearly formed after the peak metamorphism.  $\text{H}_2\text{O}$  apparently penetrated the KSZ during uplift and may be associated with minor retrograde metamorphism (which is manifested by sericitized feldspars and epidote-chlorite alteration on some of the mafic mineral phases). The mixed  $\text{H}_2\text{O}-\text{CO}_2$  inclusions form where a trail of  $\text{H}_2\text{O}$  crosses a trail of earlier  $\text{CO}_2$ .

#### Interpretation of Fluid Inclusion Data

Several inferences can be made from the above data.  $\text{CO}_2$  appears to have been the fluid phase present during the peak metamorphism. Small amounts of  $\text{CH}_4$ , present in the metasedimentary units, may have been locally derived. Two rocks, a tonalite dike and an amphibolite, possess high density  $\text{CO}_2$  inclusions which, in the case of the tonalite dike, were trapped during crystallization of the host quartz. The corresponding isochore for these dense inclusions passes through the lower portion of the estimated P-T conditions of metamorphism of the KSZ. After entrapment of these high density  $\text{CO}_2$  inclusions, the P-T path of the KSZ granulites is constrained to have remained within 1.5 kbar of the  $1.05 \text{ g/cm}^3$  isochore (the shaded region in Fig. 4A). If the rocks passed below this range, the fluid inclusions would have decrepitated due to the large pressure differential thus created between the interior and exterior of the fluid inclusion (Hollister et al., 1979). Therefore the KSZ was not uplifted while retaining high temperatures, as the Tertiary coast range granulites of British Columbia (path B, Fig. 4) (Hollister, 1979). Additionally, the KSZ granulites could not have cooled isobarically, producing denser, late-stage inclusions, as Swanenberg (1980) found for Precambrian granulites from southern Norway (path C, Fig. 4); the morphologically later inclusions in the KSZ invariably have lower densities. The KSZ granulites were uplifted along the path shown in Fig. 4A. As the P and T dropped,  $\text{CO}_2$  was released and re-trapped, forming the lower density inclusions. The lowest density  $\text{CO}_2$  inclusions present are  $0.5 \text{ g/cm}^3$ , and must have been trapped along that isochore within the shaded region (Fig. 4A).  $\text{H}_2\text{O}$  penetrated the KSZ as higher levels were reached (producing the retrograde assemblages present in some units) and was trapped, as fractures in the quartz continued to form and heal.

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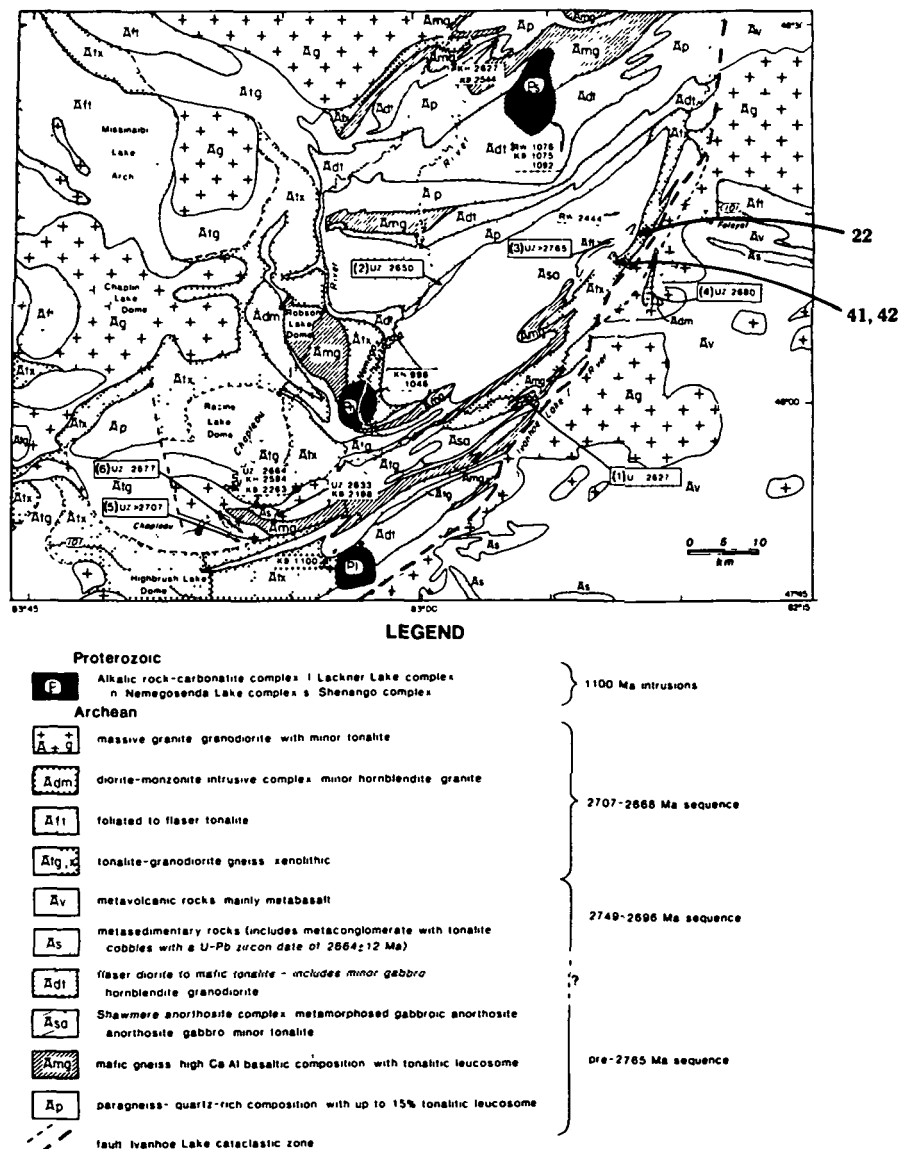


Fig. 1, Geologic map of the Kapuskasing structural zone showing sample localities for this study (from Percival and Krogh, 1983).

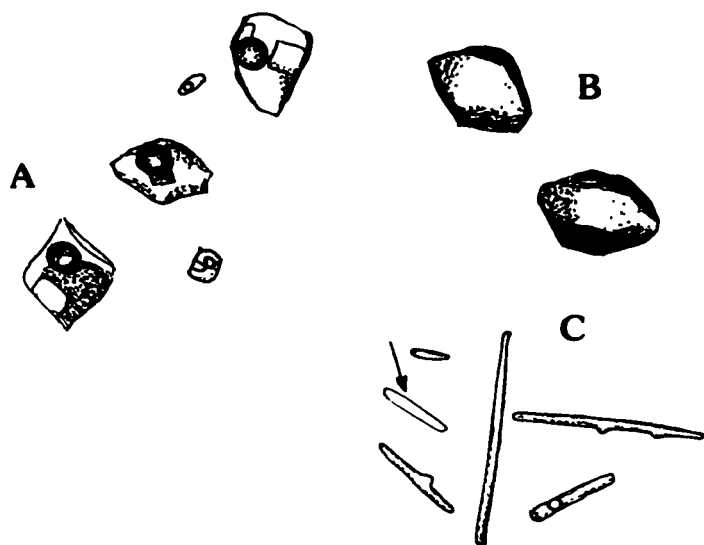


Figure 2, A: morphology of  $\text{H}_2\text{O-NaCl}$  fluid inclusions, B: negative crystal form  $\text{CO}_2$  fluid inclusions, C: acicular  $\text{CO}_2$  fluid inclusions, arrow points to acicular carbonate grain.

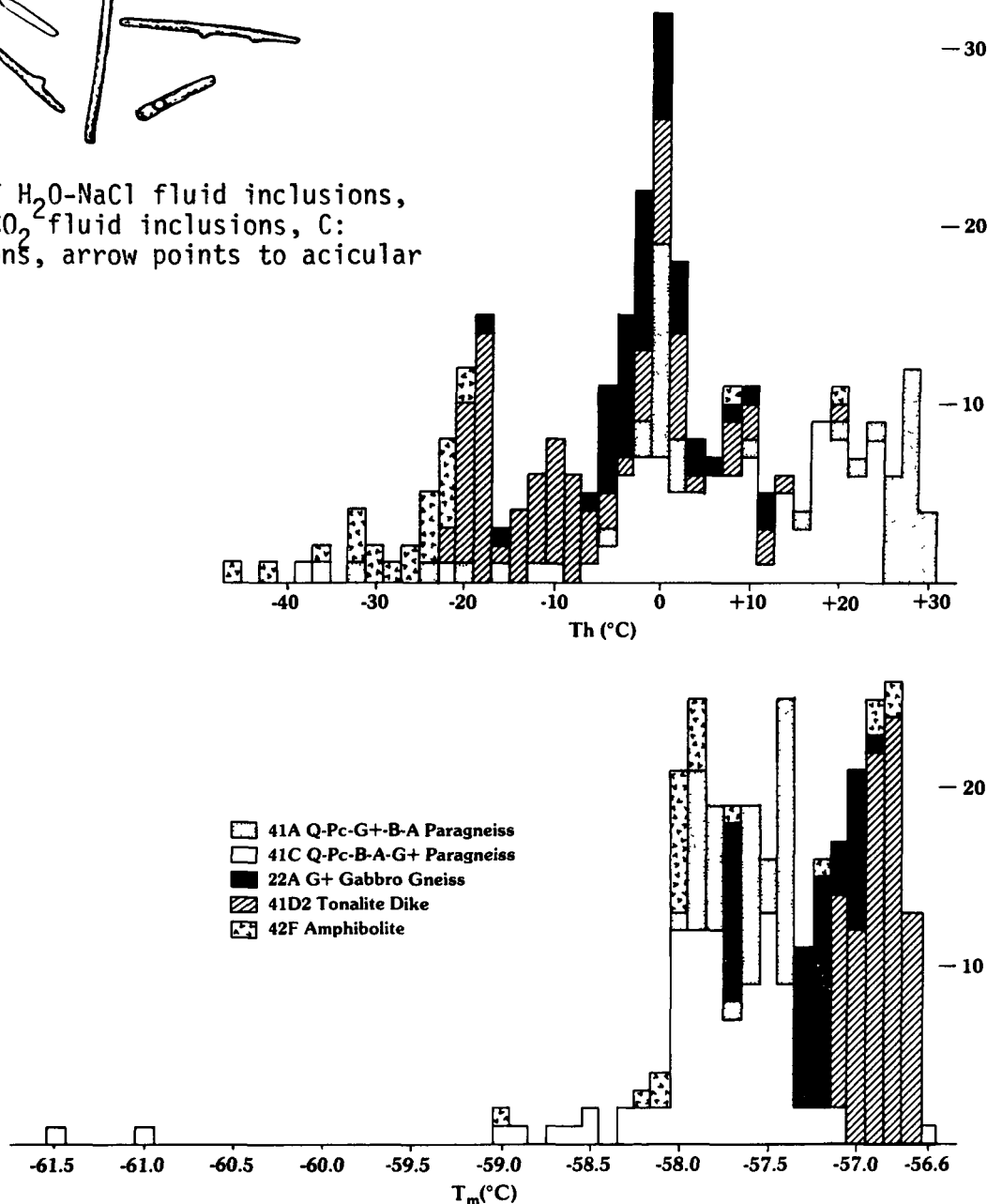


Fig. 3, Temperature of homogenization ( $T_h$ ) and temperature of melting ( $T_m$ ) for  $\text{CO}_2$ -rich fluid inclusions from high-grade rocks from the Kapuskasing structural zone, Ontario. Q = quartz, Pc = plagioclase, Gt = garnet, B = biotite, A = amphibole.

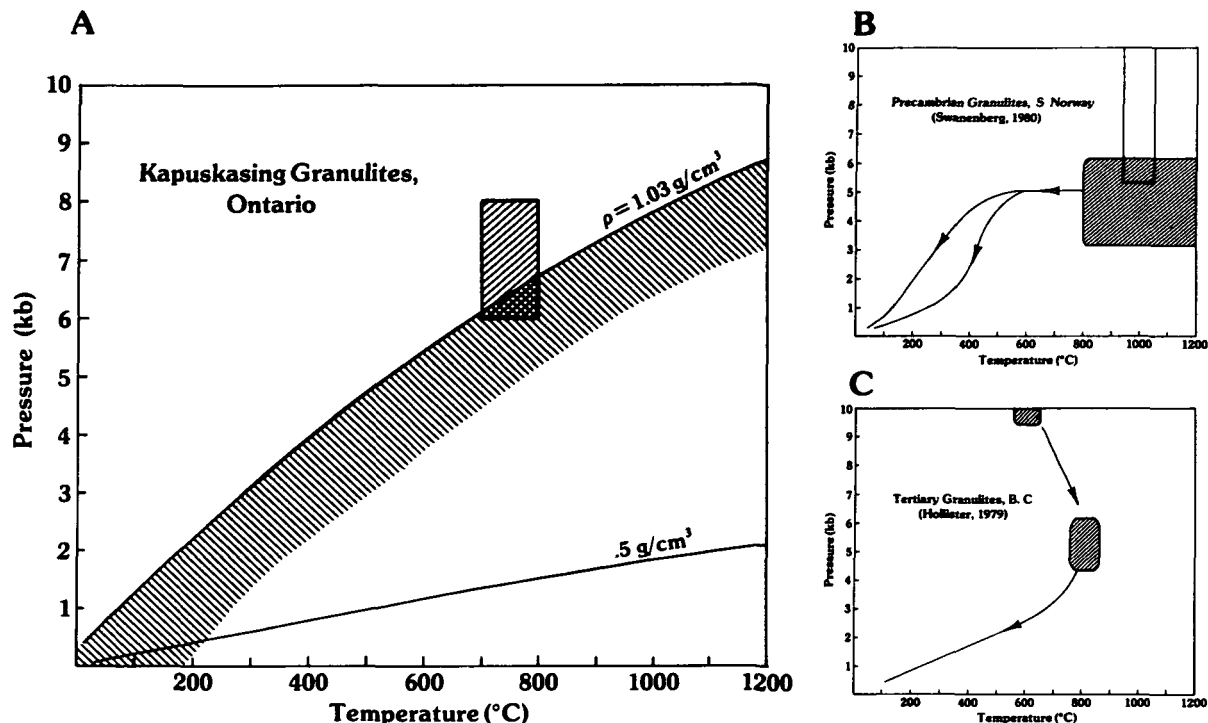


Fig. 4, A: P and T of metamorphism of the KSZ (from Percival and Card, 1933) and uplift path deduced by high density fluid inclusions ( $1.03 \text{ g/cm}^3$ ). Isochore with  $\rho = .5 \text{ g/cm}^3$  represents least dense  $\text{CO}_2$  inclusions in KSZ. These lower density inclusions must have been trapped along the  $.5 \text{ g/cm}^3$  isochore within the shaded region. B: conditions of metamorphism and uplift path for S. Norway granulites from Swanenberg (1980). C: conditions of metamorphism and uplift path for Tertiary Granulites from the Coast Range, British Columbia from Hollister, 1979.

GEOCHEMISTRY OF VOLCANIC ROCKS FROM THE WAWA GREENSTONE BELT, ONTARIO, Klaus J. Schulz, U.S. Geological Survey, Reston, VA 22092, Paul J. Sylvester, Dept. of Earth and Planetary Sciences, Washington University, St. Louis, MO 63130, and K. Attoh, Dept. of Geology, Hope College, Holland, MI 49423

The Wawa greenstone belt is located in the District of Algoma and extends east-northeast from Lake Superior to the western part of the Sudbury District in Ontario, Canada. Recent mapping by Attoh (1,2) has shown that an unconformity at the base of the Doré Formation and equivalent sedimentary rocks marks a significant stratigraphic break which can be traced throughout the volcanic belt. This break has been used to subdivide the volcanic-sedimentary succession into pre- and post-Doré sequences (2). The pre-Doré sequence includes at least two cycles of mafic-to-felsic volcanism, each capped by an iron-formation unit. The post-Doré sequence includes an older mafic-to-felsic unit, which directly overlies sedimentary rocks correlated with the Doré Formation, and a younger felsic breccia unit interpreted to have formed as debris flows from a felsic volcanic center (2).

In the present study, samples from both the pre- and post-Doré volcanic sequences were analyzed for major and trace elements, including rare earths (REE). This preliminary study is part of an ongoing program to assess the petrogenesis of the volcanic rocks of the Wawa greenstone belt.

#### Pre-Doré Volcanic Rocks

Two volcanic suites have been recognized in the pre-Doré volcanic sequence. The first is represented by tholeiitic basalts which apparently make up only a small part of the sequence. These are similar in both major and trace elements to typical Archean tholeiitic basalts such as those of the Ely Greenstone in the Vermilion District of Minnesota (3,4). The other and more abundant suite is calc-alkaline and represented by basaltic andesites, dacites and rhyolites. Andesites (i.e.  $\text{SiO}_2 = 54\text{--}62$  wt.%) appear to be largely absent.

The basaltic andesites have high  $\text{Al}_2\text{O}_3$  contents, enriched light REE ( $[\text{La}/\text{Sm}]_N = 1.9\text{--}3.3$ ), lower contents of Sc and Co, and higher contents of Hf, Zr, Ta, and Th compared with the tholeiitic basalts.

Of the three dacites analyzed, one has REE abundances similar to the basaltic andesites except for a small negative Eu anomaly, and has Sc, Co, Cr, Ta, Th, Hf, and Zr values intermediate between the basaltic andesites and rhyolites. The other two dacites, sampled from the area of Michipicoten Harbour, are distinctive in having highly depleted heavy REE abundances and steep REE patterns ( $[\text{La}/\text{Yb}]_N = 22\text{--}23$ ).

The rhyolites are generally similar in composition (except one sample from west of Andre Lake and two samples from the Michipicoten Harbour area), having enriched light REE ( $[\text{La}/\text{Sm}]_N = 3.8\text{--}4.2$ ), negative Eu anomalies, and low compatible element contents. These rhyolites tend to lie along the compositional trends defined by the basaltic andesite and dacite samples implying that they may be the product of fractional crystallization from the same parent magma. However, the apparent lack of andesites, if substantiated, may require a more complex model for the petrogenesis of these rhyolites. The two rhyolites from the Michipicoten Harbour area, when compared with the major rhyolite group, have higher REE abundances with larger negative Eu anomalies, markedly higher heavy REE contents, and are also enriched in other incompatible elements. The rhyolite from west of Andre Lake has lower REE, Hf, Zr, Ta, Th, Sc and Co abundances than the major rhyolite group and shows no Eu anomaly.



## Post-Doré Volcanic Rocks

Tholeiitic and calc-alkaline suites are both present in the post-Doré volcanic sequence. The tholeiitic suite forms part of the older post-Doré volcanic unit and is represented solely by basalts which are compositionally similar to those of the lower sequence. The rest of this older unit is composed of calc-alkaline rhyolites which can be divided into low SiO<sub>2</sub> (71-74 wt.%) and high SiO<sub>2</sub> (78-79 wt.%) groups. These rhyolites have steeper REE patterns than the lower volcanic unit rhyolites. The low SiO<sub>2</sub> rhyolites have no or small Eu anomalies, whereas the high SiO<sub>2</sub> rhyolites have prominent negative Eu anomalies, lower Sr (<90 ppm), and variable light REE abundances (La=20-48 ppm). The high SiO<sub>2</sub> rhyolites probably represent fractionated melts of the low SiO<sub>2</sub> rhyolite group.

Two samples analyzed from the younger felsic breccia unit are dacitic in composition, have high Sr contents (305 and 477 ppm), steep REE patterns ([La/Yb]<sub>N</sub>=10.7 and 21.7) with [La]<sub>N</sub>=74 and 204, and small negative Eu anomalies.

## Discussion

The present data show that although the tholeiitic basalts of the pre- and post-Doré volcanic sequences are compositionally similar, the calc-alkaline suites are not. Although andesites appear to be largely absent from the pre-Doré calc-alkaline suite, the general compositional continuity of the basaltic andesites, dacites and rhyolites suggests some co-genetic relationship.

The rhyolites of the older post-Doré volcanic sequence are compositionally distinct, particularly in terms of their REE patterns, from the pre-Doré rhyolites (i.e., post-Doré rhyolites have lower heavy REE abundances, steeper REE slopes and the low SiO<sub>2</sub> rhyolites lack Eu anomalies). The apparent lack of more mafic calc-alkaline rocks in this post-Doré sequence makes a fractional crystallization origin for these rhyolites unlikely and suggests they may be a product of partial melting.

Campbell and others (5) have shown that Archean felsic volcanic rocks associated with Cu-Zn massive sulfide mineralization have flat REE patterns with well-developed Eu anomalies whereas rhyolites from barren volcanic sequences have steeper REE patterns with weak or absent Eu anomalies. The rhyolites of the Wawa greenstone belt (both pre- and post-Doré) appear to have the general REE characteristics of the latter group, which is consistent with the lack of known massive sulfide mineralization in the Wawa belt.

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**GRANITIC ROCKS AND METASEDIMENTS IN ARCHEAN CRUST, RAINY LAKE AREA, ONTARIO: ND ISOTOPE EVIDENCE FOR MANTLE-LIKE SM/ND SOURCES, S.B. Shirey and G.N. Hanson, Earth and Space Sciences, SUNY, Stony Brook, NY 11794**

Granitoids, felsic volcanic rocks and clastic metasediments are typical rocks in Archean granite-greenstone belts that could have formed from pre-existing continental crust. Understanding the petrogenesis of such rocks is important to assess the relative roles of new crust formation or old crust recycling in the formation of granite-greenstone belts.

The Rainy Lake area, Ontario, is a 35 by 10 km fault-bounded, granite greenstone terrane located between the Wabigoon metavolcanic-plutonic superbelt and the Quetico gneiss-metasediment superbelt of the Superior Province. Its structure and composition are similar to the granite-greenstone belts which surround the large oval complexes of polycyclic granite and granite gneiss in the Wabigoon superbelt.

The stratigraphic and structural interpretations (1,2,3,4,5,6) of rocks in the area yield the following sequence (oldest to youngest): interbedded basalts and rhyolites, graywackes, gabbros to anorthosites, tonalites, syenodiorites and granodiorites. The early supracrustal units display recumbent, near-isoclinal folds and metamorphism into the amphibolite grade has affected all but the granodiorites. U-Pb isotope ratios on zircons and sphene (7,8,9) from all the rocks except the basalts and gabbros fall about a chord with an upper intercept of  $2670 \pm 30$  My, representing a mean age for the belt.

The major element chemistry of the rocks (Fig. 1) and REE chemistry (Fig. 2) are diverse, suggesting each group of rocks was probably derived by a separate process. For example, two types of mantle sources, one strongly light REE enriched and the other only slightly light REE enriched are implied by the REE distribution in the basalts and syenodiorites. The low heavy REE content of the graywackes suggests the existence of a heavy REE depleted crustal material that was sampled during erosion and deposition. The high heavy REE content of the rhyolites rules out the basalt as a precursor for any reasonable extent of melting or fractional crystallization. Sampling of different source areas by melting or erosion would be expected to yield rocks with a variety of Nd isotope initial ratios that would be a function of the age of each source and its time-integrated Sm/Nd.

$^{143}\text{Nd}/^{144}\text{Nd}$  from the suite of granitoids, rhyolites and metasediments fit within error about a line whose slope corresponds to a mean age of 2670 My with the exception of two samples (Fig. 3). This suggests metamorphic recrystallization of the rocks has not strongly affected the Sm-Nd system on the scale of whole-rock samples at a time much later than the formation of the belt. Because the relative ages of each rock group is well known from field relations, a relative model initial  $^{143}\text{Nd}/^{144}\text{Nd}$  can be calculated for each of the rocks and plotted on its growth curve in epsilon Nd versus time (Fig. 4).

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Nd isotope ratios on mafic-ultramafic volcanic rocks (10,11,12) suggest the worldwide Archean mantle at 2700 My had a composition that ranged from +4.0 to -2.5 epsilon Nd units. Early in the development of the Rainy Lake belt, the mantle beneath it must have had a range in composition capable of yielding the basalts, gabbros and anorthosites presently exposed. Nd isotope data on these rock types (13,14) suggest the local mantle had a restricted range of -0.1 to +1.9 epsilon Nd units (Fig. 4). Nd isotope growth curves for metasediments, rhyolites and granitoids project through the values for the local mantle within error and show an identical range of 2 epsilon units at 2700 My. The source areas for these rocks could contain significantly older ultramafic-mafic rocks with Sm/Nd similar to CHUR. This possibility can not be resolved with Nd isotope data. However the data do rule out the possibility that the granitoids were derived by melting of

significantly older rocks with Sm/Nd ratios similar to typical continental crust. One likely possibility is that the rocks in the Rainy Lake area had a short-lived history in the crust and the precursors to these rocks ultimately removed from the mantle shortly before the formation of the belt.

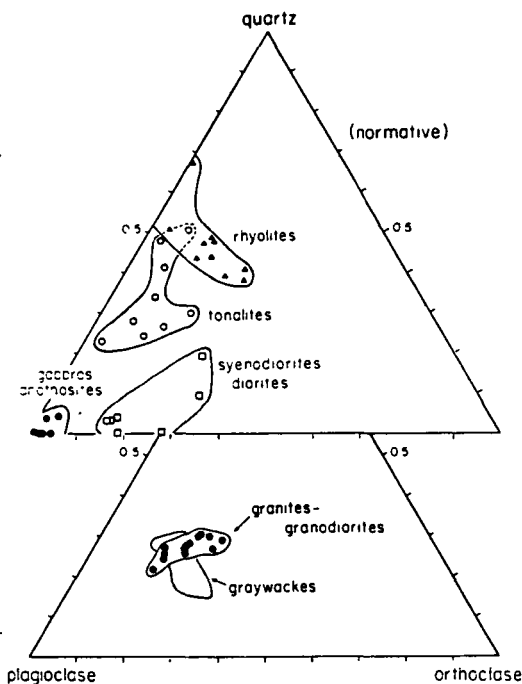
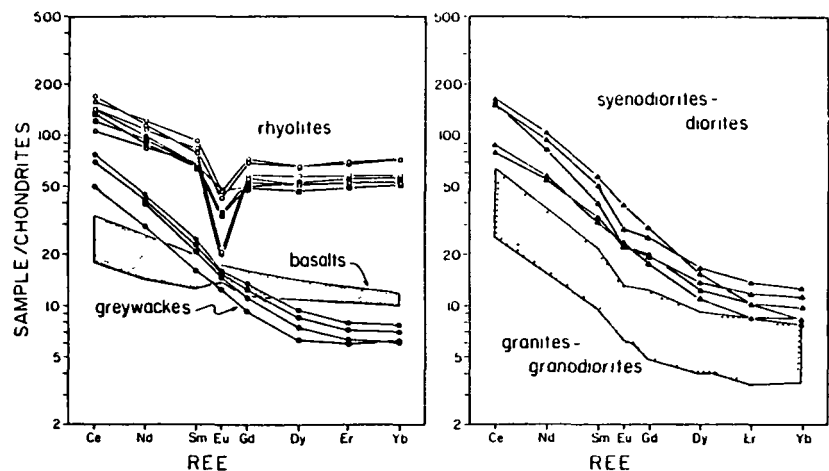


Figure 1.

Figure 2.



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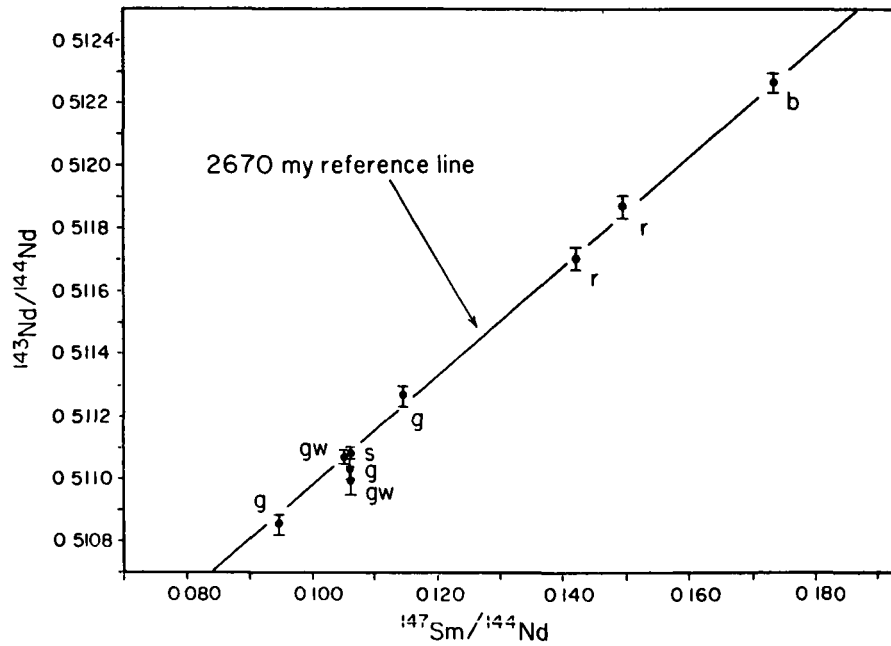
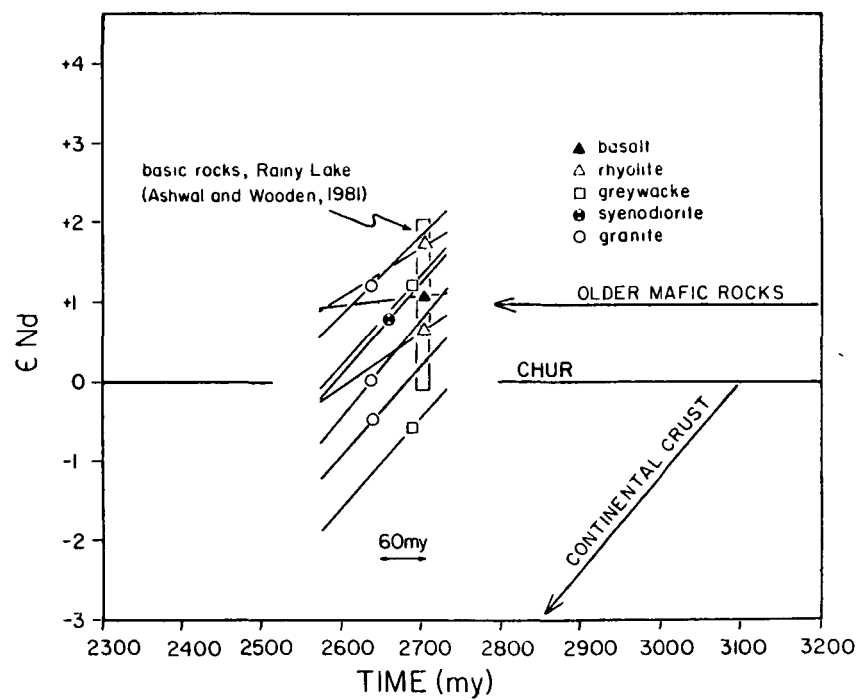


Figure 3.

Figure 4.



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## CONTINENTAL CRUSTAL COMPOSITION AND LOWER CRUSTAL MODELS.

S.R. Taylor, Research School of Earth Sciences, Australian National University, Canberra, Australia.

The composition of the upper crust is well established as being close to that of granodiorite (Table 1, col A). The upper crustal composition is reflected in the uniform REE abundances in shales<sup>1</sup> which represent an homogenisation of the various igneous REE patterns (Fig. 1). This composition can only persist to depths of 10-15 km, for heat flow and geochemical balance reasons. The composition of the total crust is model dependent. One constraint is that it should be capable of generating the upper granodioritic (S.L.) crust by partial melting within the crust. One proposed composition is given in Table 1 Col. B. This composition is based on the "andesite" model, which assumes that the total crust has grown by accretion of island arc material. However, the relationship between the generation of island arc magmas and subduction zones implies that a plate tectonic regime was operative during the formation of the bulk of the continental crust. The evidence for such processes does not extend clearly beyond the late Proterozoic, by which time perhaps 80% of continental growth had probably been accomplished. Figure 3 shows a representation of the growth rate of the continental crust<sup>1</sup>. It can be noted that freeboard constraints, reflecting essential constancy of continental and oceanic volumes, clearly apply, in this type of model, well back into the Proterozoic, but are not necessarily valid in the Archean<sup>2</sup>.

Archean upper crustal compositions, derived from REE sedimentary rock patterns, show a more basic upper crust than occurs in Post-Archean time<sup>3</sup>. The data are consistent with an upper crust derived from the bimodal basaltic - felsic Archean igneous suites. The bulk composition of the Archean crust appears to be only slightly more basic than the upper crust. There is only minor evidence of intracrustal melting and the production of K-rich granites with Eu depletion must comprise less than 10% of the exposed upper crust, from the sedimentary REE data, which only very rarely show Eu depletion in contrast to Post-Archean shales. On the model adopted here, the bulk of the crust has grown by 2.5 Ae and hence the bulk compositions may reflect that of the Archean bimodal basaltic - (tonalite-trondhjemite) suite. This is not very different in composition to that of the "andesite" model, except that it contains more Ni and Cr.

The composition of the lower crust, which comprises 60-80% of the continental crust, remains a major unknown factor for models of terrestrial crustal evolution. For the lower crust, we lack those large scale natural sampling processes (such as production of clastic sediments or loess) which have simplified the task of arriving at upper crustal compositions. Lower crustal samples are either random (as xenoliths in volcanic pipes) or from restricted outcrop areas of granulite terrains. The lower crust is almost certainly heterogeneous in detail, and may be further complicated by the presence of imbricate thrust sheets. One constraint is that the granodioritic (S.L.) rocks of the upper crust originated by partial melting within the crust, at depths of less than 40 km<sup>4,5</sup>. The lower crust must accordingly include many regions from which granitic melts (S.L.) have been extracted. If we recognize such material in the scattered samples available it will provide valuable limitations on the bulk composition of the crust.

Various approaches are possible. One is to model the bulk crust and calculate residual compositions following the extraction of granitic melts.

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In Table 1 (cols. D,E), such calculations are presented for the extraction of a minimum granitic melt from proposed total crustal compositions allowing for 10 and 20% extraction of melt. A slightly different approach is to extract the known upper crustal composition (Col. A) from the model total crust, assuming that the upper crust forms 33% (Col. F) and 20% (Col. G) of the total. These four compositions indicate that the lower crust should contain compositions high in  $\text{Al}_2\text{O}_3$ ,  $\text{CaO}$ , low in  $\text{K}_2\text{O}$ , with positive Eu anomalies ( $\text{Eu}/\text{Eu}^* > 1$ ) and  $\text{Nd}_\text{N}/\text{Sm}_\text{N}$  ratios approaching chondritic values. Figs. 1 and 2 show the upper crustal REE patterns, and the predicted REE patterns for the lower crust.

A second approach is to examine the composition of dry granulite samples, formed at lower crustal temperatures and pressures to see whether they match the model calculations. Fyfe<sup>5</sup> has pointed out the importance of removal of  $\text{H}_2\text{O}$  and minimum granitic melts at the upper amphibolite grade of metamorphism, allowing the development of the anhydrous mineralogy typical of the granulite facies. Granulite facies rocks can be expected to show wide variations in composition due to several processes:

- (A) Development of granulite facies mineralogy in dry source rocks from which a granitic melt has been extracted during amphibolite facies metamorphism<sup>5</sup>.
- (B) Dehydration of source rocks with loss of an hydrous fluid phase, resulting in granulites depleted in alkalis and U<sup>6</sup>.
- (C) Dehydration without partial melting or loss of trace elements (eg Jequie complex, Brazil)<sup>7</sup>.
- (D) Dehydration accompanied by loss of  $\text{CO}_2$  (eg Southern India)<sup>8</sup>
- (E) Subsequent retrograde metamorphism to produce amphibolite facies mineralogy in which any or all of the above processes have operated.

Accordingly, much complexity is expected, and shown by the random examples of lower crustal compositions available. In Table 1, Cols. H to Q, data are given for a suite of Lewisian<sup>9</sup> and Scourian granulites<sup>10</sup>, granulite xenoliths from Lesotho<sup>11,12</sup> Bournac, France<sup>13</sup> and eclogites from Sauviat-sur-Vige, France<sup>14</sup>. These compositions are typified by high  $\text{Al}_2\text{O}_3$ , and  $\text{CaO}$ , low  $\text{K}_2\text{O}$  and positive europium anomalies. Fig. 4 shows the REE patterns. Figure 5 shows similar REE patterns in granulite xenoliths from alkali basalts from Central Hoggar, Algeria<sup>15</sup>. All these patterns display Eu enrichment, which accordingly is not uncommon in lower crustal material, although it is rare in upper crustal rocks. The major and trace element compositions tend to show much variation, as noted by Griffin et al<sup>11</sup> in their study of the Lesotho xenoliths. In this example, minerals such as garnet show Eu enrichment and the development of the bulk rock REE pattern, with positive Eu anomalies, clearly predates the granulite facies metamorphism. Accordingly, the extraction of granitic melts prior to granulite facies metamorphism will change the bulk rock composition, including development of the Eu enrichment (since the granitic melts are typified by Eu depletion). It should be noted that  $\text{Nd}/\text{Sm}$  ratios (Table 1) are lower than either upper crust or total crustal estimates, placing important constraints on isotopic models of the lower crust.

Condie et al<sup>16</sup> report REE patterns with strong Eu enrichment from Archean tonalitic gneisses in Southern India. Minor amounts of granitic gneisses and tonalites have probably developed from the tonalites. The origin of the Eu enrichment in the tonalites is ascribed to partial melting of a mafic source enriched in Eu. An alternative hypothesis, presented here, is that the present tonalite chemistry is residual and that the Eu enrichment

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has been generated by extraction of an Eu depleted granodioritic melt. Fig. 6 shows that a mixture of upper crust and the Southern Indian tonalites generates REE patterns which resemble normal Archean tonalites, with no Eu anomaly. The other trace element data are also consistent with the proposal that the Southern Indian tonalites are residual from earlier tonalitic parents. The REE patterns show a close resemblance to the Scourian data. The well studied Scourian succession has been the subject of varying interpretations. Pride and Muecke<sup>10</sup> note the following arguments in favour of extraction of partial melts (a) anhydrous nature of the complex (b) incompatible element depletion (c) narrow range of mineral compositions (d) major element trends unlike those of upper crustal igneous rock sequences (e) REE abundances are lower than those typical of upper crustal rocks, with enrichment of europium. Accordingly, there is much evidence for europium enrichment in lower crustal samples. Whether this is caused by melt extraction leaving Eu in residual plagioclase remains to be fully tested.

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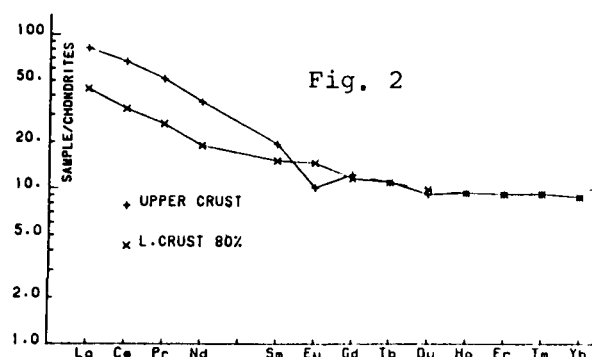
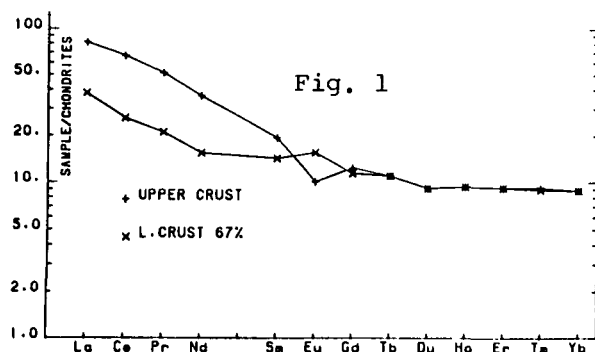


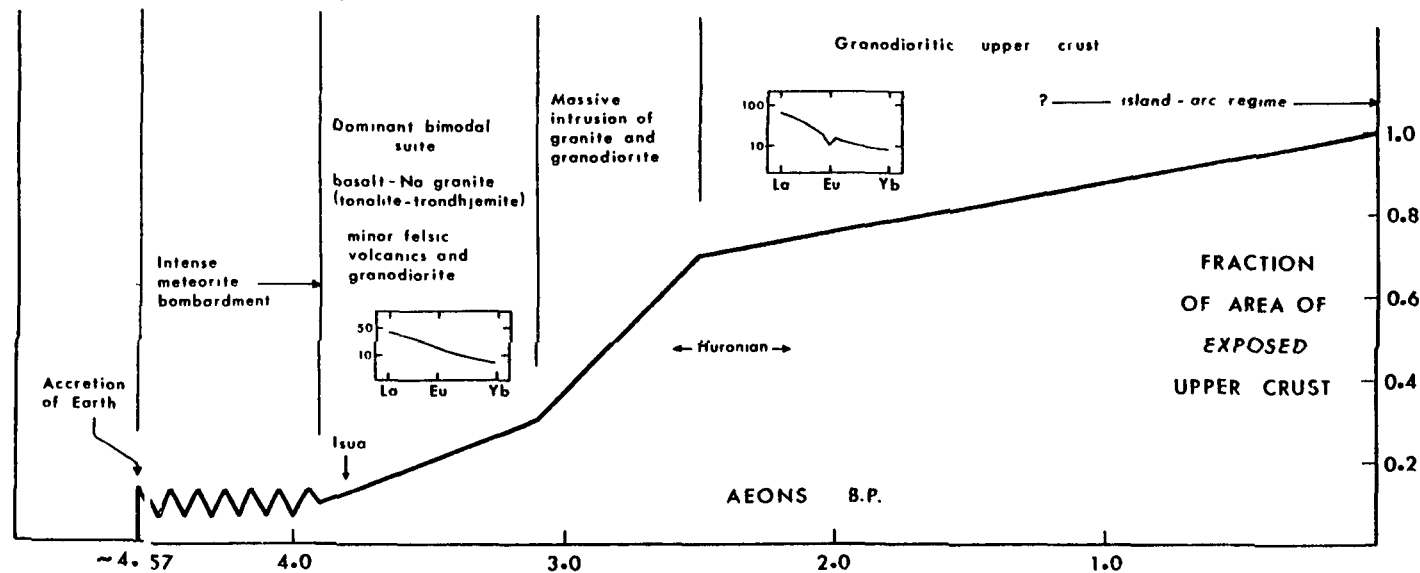


Table 1

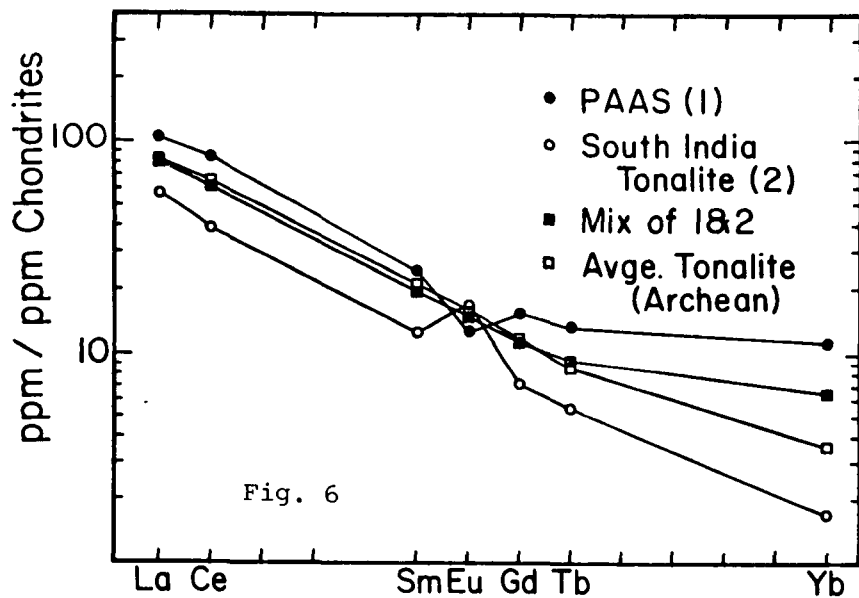
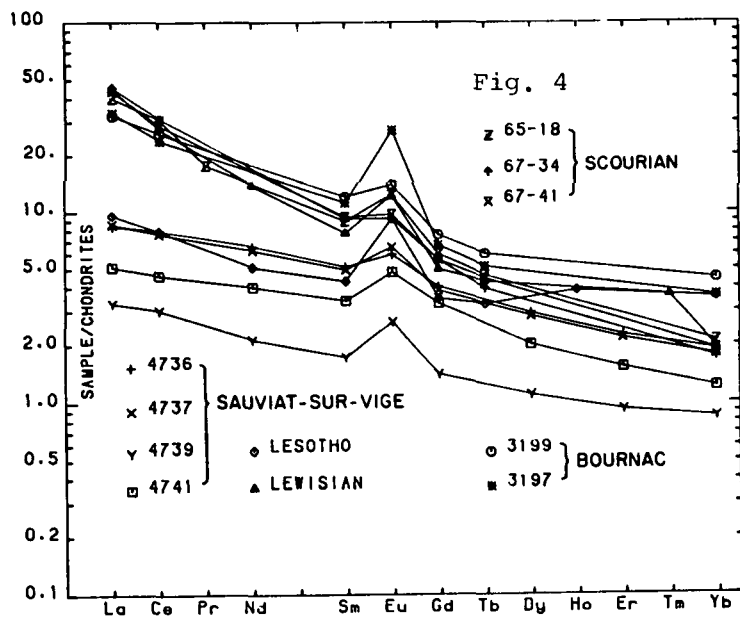
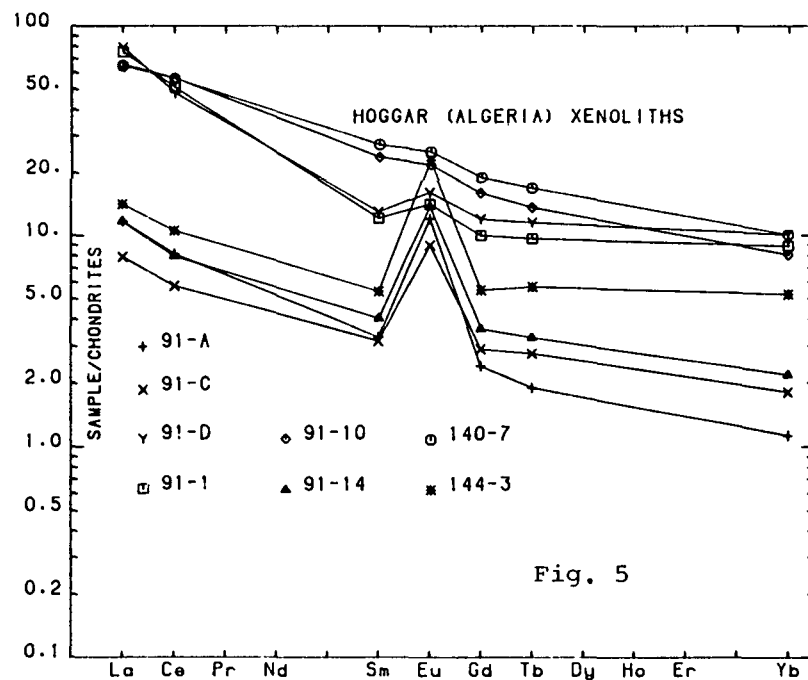
	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O	P	Q
SiO <sub>2</sub>	66.0	58.0	76.5	53.4	55.9	54.0	56.0	61.3	50.6	52.2	52.6	57.5	66.5	65.3	49.5	48.5	41.9
TiO <sub>2</sub>	0.6	0.8	0.1	1.0	0.9	0.9	0.9	0.56	0.77	1.3	1.2	0.96	0.54	0.51	0.70	0.55	0.23
Al <sub>2</sub> O <sub>3</sub>	16.0	18.0	12.9	19.3	18.6	19.0	18.5	18.1	17.9	19.5	24.5	19.5	16.5	16.2	18.5	19.7	18.0
FeO <sub>T</sub>	4.5	7.5	1.1	9.1	8.2	9.0	8.3	5.4	8.8	9.2	5.9	5.7	4.0	4.2	6.6	5.5	13.3
MgO	2.3	3.5	0.1	4.4	3.9	4.1	3.8	2.4	8.2	5.9	2.0	2.8	2.0	2.3	8.3	8.7	27.6
CaO	3.5	7.5	0.7	9.2	8.3	9.5	8.5	5.1	9.4	8.7	7.9	6.6	5.4	5.3	14.5	14.8	5.3
Na <sub>2</sub> O	3.8	3.5	3.8	3.4	3.5	3.4	3.4	4.9	3.3	2.7	4.0	4.9	4.1	4.3	1.6	2.0	0.65
K <sub>2</sub> O	3.3	1.5	4.7	0.7	1.1	0.6	1.1	2.4	1.0	0.55	1.7	0.9	0.8	0.9	0.4	0.26	-
Eu/Eu*	0.64	1.0	0.45	1.35	1.12	1.22	1.10	1.97	2.4	1.46	3.14	1.69	1.29	1.34	1.31	1.50	1.70
Nd <sub>N</sub> /Sm <sub>N</sub>	1.88	1.41	1.72	1.14	1.29	1.09	1.26	1.79	1.24	-	-	-	-	-	1.29	1.26	1.24

A Upper Crust<sup>1</sup> B Total Crust<sup>1</sup> C Minimum melt composition<sup>17</sup> D Residue following 20% melt extraction from total crust  
 E Residue following 10% melt extraction F Predicted lower crust composition after extracting 33% upper crust Col. 1 from Col. 2 G Predicted lower crust composition following extraction of 20% upper crust.  
 H Lewisian granulites, Ave of 8<sup>2</sup> I Lesotho garnet granulites PHN 1670, 2533, 2852 and L.13<sup>11 12</sup>  
 J Granulite xenolith, Bournac 3199<sup>13</sup> K Granulite xenolith, Bournac 3197<sup>13</sup> L Scourian granulite 65-16<sup>10</sup>  
 M Ibid, 67-34<sup>10</sup> N Ibid, 67-41<sup>10</sup> O Sauviat-Sur-Vige metagabbro 4736<sup>14</sup> P Ibid 4737 eclogite<sup>14</sup>  
 Q Ibid 4739 garnet peridotite<sup>14</sup>

Fig. 3



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ARCHEAN CRUST-MANTLE GEOCHEMICAL DIFFERENTIATION, G.R. Tilton,  
Geological Sciences, University of California, Santa Barbara, CA 93106

Isotope measurements on carbonatite complexes and komatiites can provide information on the geochemical character and geochemical evolution of the mantle, including sub-continental mantle. Measurements on young samples establish the validity of the method. These are based on Sr, Nd and Pb data from the Tertiary-Mesozoic Gorgona komatiite (1, 2) and Sr and Pb data from the Cretaceous Oka carbonatite complex (3, 4). In both cases the data describe a LIL element-depleted source similar to that observed presently in MORB.

Carbonatite data have been used to study the mantle beneath the Superior Province of the Canadian Shield one billion years (1 AE) ago. The framework for this investigation was established by Bell et al. (3) who showed that large areas of the province appear to be underlain by LIL element-depleted mantle ( $^{87}\text{Sr}/^{86}\text{Sr} = 0.7028$ ) at 1 AE ago. Additionally Bell et al. found four complexes to have higher initial Sr ratios ( $^{87}\text{Sr}/^{86}\text{Sr} = 0.7038$ ), which they correlated with less depleted (bulk earth?) mantle sources, or possibly crustal contamination.

We have determined Pb isotope relationships in four of the complexes studied by Bell et al. In favorable cases the carbonates from the complexes yield negligible in situ radiogenic Pb corrections ( $\mu = 0.01 - 2$ ), allowing accurate determination of initial ratios. Initial ratios for six samples from three of the complexes with 0.7028 initial Sr ratios (Firesand, Prairie Lake, Killala) plot along a regression line given by  $^{207}\text{Pb}/^{204}\text{Pb} = 0.128$   $^{206}\text{Pb}/^{204}\text{Pb} + 13.186$ , with  $^{206}\text{Pb}/^{204}\text{Pb}$  varying from 16.48 to 17.08. The data plot distinctly below crustal Pb evolution curves as given, for example, by Stacey and Kramers (5). The slope of the regression line, 0.128, differs significantly from the value expected from contamination with 1.0 AE Pb (0.0725) or 2.7 AE Pb (0.185). The carbonatite regression line plots to the left of the modern MORB regression line and has a slightly greater slope, apparently describing Pb isotope relationships for a billion-year-old MORB-like mantle source. The Pb and Sr data agree in suggesting that LIL element-depleted mantle existed beneath large areas of the Superior Province one billion years ago as far inland as the present-day Lake Superior region. Pb data from a fourth complex (Lake Nemegosenda,  $^{87}\text{Sr}/^{86}\text{Sr} = 0.7038$ ) plot above the crustal Pb evolution curve, and agree with the Sr data in indicating either crustal contamination, or origin in more LIL-enriched mantle.

In contrast to the above results, isotopic data from 2.7 AE rocks of the Superior Province suggest that depleted mantle was of limited extent, and not sufficiently aged to have acquired an isotopic signature at that time. Sr data from the alkaline complex at Poohbah Lake (3) plotted nearly in the "bulk earth" field in a Sr evolution diagram ( $^{87}\text{Sr}/^{86}\text{Sr} = 0.7012$ ) rather than below the field as in the case of the Oka and most of the billion year old complexes. Zindler et al. (6) showed that initial  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios in 2.7 AE komatiites and tholeiites of Munro Township, Albitibi District plot on the chondrite evolution curve at 2.7 AE, rather than above the curve as is observed for rocks from depleted mantle, e.g., MORB.

More recently, a Pb isotope study has been started on rocks from Munro Township (2). Sampling to date includes komatiite, tholeiite and sulfide ores. The komatiites are taken to represent mantle isotope relations, while the ores should characterize crustal isotopic compositions. When the data are plotted in a  $^{207}\text{Pb}/^{204}\text{Pb} - ^{206}\text{Pb}/^{204}\text{Pb}$  diagram (Fig. 1) they closely fit an isochron giving an age of 2.65 AE. Statistical analysis of the regression line yields a Y axis intercept of  $12.101 \pm 0.035$ , a deviation within the  $2\sigma$

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errors of the individual  $^{207}\text{Pb}/^{204}\text{Pb}$  ratio measurements. This agreement is taken to indicate that all members of the suite have nearly identical age and initial Pb isotope ratios. The Pb data thus appear to be consistent with the SR data from Poohbah Lake and Nd data from Munro Township in failing to identify a LIL depleted mantle source for the komatiites and tholeiites 2.7 AE ago. An analogous case for Pb isotopic data from the Fennoscandian Shield of Finland was given by Vidal et al. (7), with the exception that the Finland regression line is systematically displaced above the Munro Township line, as shown in Fig. 1. The Finland suite includes granitic rocks that plot along the isochron with the komatiites and tholeiites. The only granitic Pb isotope data from the Abitibi District available so far are given in an abstract by Gariepy et al. (8), who report "large" variations in  $^{207}\text{Pb}/^{204}\text{Pb}$  ratios in unmetamorphosed plutons. The contrast between the granite and ore data may indicate that the ores average out differences between individual plutons. Further isotopic studies of Pb are underway in the granitic rocks from Munro Township.

Although evidence for depleted mantle has been observed in Nd data in 3.5-3.8 AE rocks in other shield areas, Nd, Sr and Pb data suggest that depleted mantle originated ca. 2.7-3.0 AE ago in several areas beneath the Superior Province in the Canadian Shield. The relative abundance of depleted mantle on a world-wide basis in Archaean time remains to be definitively answered in future work.

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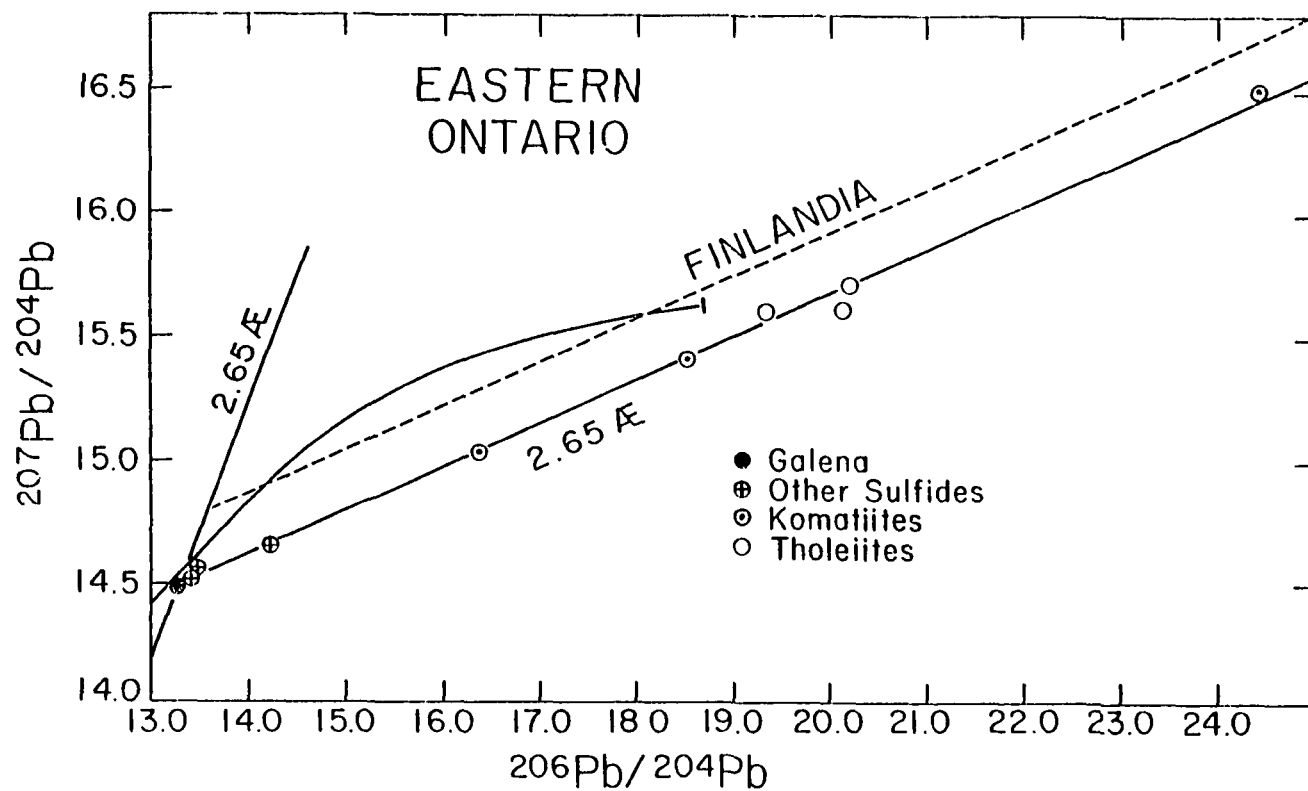


Figure 1. Pb isochron diagram for eastern Ontario samples. The dashed line is the regression line described by the Finnish data of Vidal et al. (7)

**THE PIKWITONEI GRANULITE DOMAIN: A LOWER CRUSTAL LEVEL ALONG THE CHURCHILL-SUPERIOR BOUNDARY IN CENTRAL MANITOBA.** W. Weber, Manitoba Geological Services Branch, Winnipeg, Manitoba, Canada, R3H 0W4

The greenschist to amphibolite facies tonalite-greenstone terrain of the Gods Lake subprovince grades - in a northwesterly direction - into the granulite facies Pikwitonei domain (1) at the western margins of the Superior Province,.

The transition is the result of prograde metamorphism and takes place over 50 - 100 km without any structural or lithological breaks. Locally the orthopyroxene isograd is oblique to the structural grain and transects greenstone belts, e.g., the Cross Lake belt (2).

The greenstone belts in the granulite facies and adjacent lower grade domain consist mainly of mafic and (minor) ultramafic metavolcanics, and clastic and chemical metasedimentary rocks (1,2). Typical for the greenstone belts crossed by the orthopyroxene isograd are anorthositic gabbros and anorthosites, and plagiophyric mafic flows.

Available data suggest a late Aphebian age for the prograde greenschist to granulite facies metamorphism. Peak conditions are reflected by sapphirin-bearing and opx-sillimanite quartz gneisses which indicate 10 - 11 kb pressure and temperatures of 900 - 1000°C (2).

At its western and northern edge - towards the contact with the Churchill Province - the rocks of the Pikwitonei granulite domain were overprinted by the Hudsonian orogeny; they were deformed, selectively retrogressed and recrystallized under greenschist to amphibolite facies conditions (1,2); locally they were migmatized. The Thompson belt, the Split Lake block and a linear zone south of the Fox River consists of these reworked granulites.

Proterozoic rocks of the Circum-Superior belt (3) (apparently) overlie the reworked granulites along the Fox River and in the Thompson belt. They consist of metasedimentary rocks, mafic-ultramafic metavolcanic and associated intrusive rocks; the magmatic rocks are komatiitic in nature suggesting a rifting environment (3,4).

The contact between the Superior and Churchill Province is a fault. The rocks on the Superior side of this fault are the above-mentioned reworked granulites or rocks of the Circum-Superior belt. Aphebian Kisseynew-type metasedimentary gneisses generally occur on the Churchill side (1).

The Pikwitonei granulite domain has been interpreted as to represent a lower crustal level (2,5,6) which was uplifted to the present level of erosion.

On the basis of gravimetric data this uplift has been modelled as an obduction onto the Churchill Province during the Hudsonian orogeny, similar to the Ivrea Zone (5,6). The fault between the Churchill and Superior Province has been described as suture (7).

However, field geological data suggest that the uplift of the crust is older, late Archean and/or very early Proterozoic and possibly related to a rifting event. Portions of the split-up edge of the Superior craton may be represented by the granulite grade portions of the Wollaston-Nejanilini domain (8,9).

The main evidence for an older uplift is the Molson dykes which intruded into the granulites and the adjacent lower grade terrains of the

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northwestern Superior Province when both terrains were at shallow crustal levels, and prior to the Hudsonian orogeny. Since (reworked) granulites form the basement to the Circum-Superior belt, this also suggests early uplift, prior to the deposition of the supracrustal rocks. Preliminary data, based on Pb-Pb isotopes, yield early Archean ages for the intrusion of the Molson dykes and the extrusion of the komatiites in the Thompson belt (10).

The present juxtaposition of the Superior and Churchill Province rocks is the result of a collision caused by a northward movement of the brittle Superior Province with respect to the Slave Province. Most of the deformation pattern in the southeastern Churchill Province and the fault pattern in the northwestern Superior block are the result of this relative movement (11).

The collision between the Superior and the Churchill Provinces led to overthrusting of Churchill Province rocks onto the Circum-Superior belt along the Fox River and a strike-slip fault between the Thompson belt and the Churchill Province. The intense deformation - related to the northwards movement of the Superior edge - in the southeastern Churchill Province, suggests a large, but presently unknown lateral displacement along this fault. Compressional stress perpendicular to the Thompson belt probably also some produced vertical displacements along the fault and along splay faults in the Thompson belt, and possibly minor obduction of the Superior Province onto the Churchill Province at Thompson (based on gravimetry). However, seismic reflection in the Thompson area (12) do not support major obduction.

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## THE GREENSTONE BELTS OF ZIMBABWE, James F. Wilson, University of Zimbabwe

Three ages of granite-greenstone terrane can be recognised within the Zimbabwe Archaean Craton. The oldest greenstone belt remnants constitute the volcano-sedimentary Sebakwian group dated at c. 3.5 Ga minimum on the evidence from various granites and gneisses. The more extensive, main greenstone belts comprise the dominantly volcanic Bulawayan Group and dominantly sedimentary Shamvaian Group. An unconformity within the Bulawayan Group allows its subdivision into the Lower and Upper Greenstones. The Lower Greenstones possibly form part of a granite-greenstone terrane about 2.9 Ga old. The widespread Upper Greenstones and the locally developed, unconformably overlying Shamvaian Group are about 2.7 Ga old. Two suites of late granites post-date the main greenstone belts. These comprise the tonalitic Sesombi Suite at c. 2.7 Ga and the more potash-rich Chilimanzi Suite at c. 2.6 Ga. The intrusion of the Great Dyke at c. 2.5 Ga marks the end of the Archaean.

Nappe tectonics were a feature of the ancient c. 3.5 Ga terrane. A complex granitic crust was present at an early stage but whether this formed a basement to the Sebakwian rocks is not clear.

The main greenstone belts (? 2.9 Ga and c. 2.7 Ga) however were laid down on granitic crust containing the remnants of earlier greenstone belts. Correlations of the major stratigraphic units of the (c. 2.7 Ga) Upper Greenstones across the craton indicate that these are the remains of a much more continuous cover. Certain mafic dyke swarms and a suite of dominantly ultramafic sills and layered intrusions, all of which predate the 2.6 Ga granites, help delineate the pre-Upper Greenstones basement. These ultramafic sills probably represent the dregs of magma chambers which fed the volcanic cover.

The present configuration of the greenstone belts, which is largely that of Upper Greenstones, can no longer be explained in terms of multiple granite intrusion and vertical tectonics.

**The Archean Crust in the Wawa-Chapleau-Timmins Region**

A field guidebook prepared for the 1983  
Archean Geochemistry-Early Crustal  
Genesis Field Conference

by

J.A. Percival, K.D. Card,  
Geological Survey of Canada,  
Ottawa, Ontario K1A 0E4

R.P. Sage, L.S. Jensen,  
Ontario Geological Survey,  
Toronto, Ontario M5S 1B3

and

L.E. Luhta,  
Resident Geologist,  
Ontario Ministry of Natural Resources,  
Timmins, Ontario P4N 2S7

## Introduction

The purpose of the trip is to examine the characteristics and interrelationships of Archean greenstone-granite and high-grade gneiss terranes of the Superior Province. A 300-km long west to east transect between Wawa and Timmins, Ontario will be used to illustrate regional-scale relationships.

Figure 1 shows the major geological features of the Superior Province and Figure 2 traces the trip route. The first day will be spent examining features of the Michipicoten belt, a dominantly metavolcanic portion of the Wawa subprovince. On day two, the contact relationships between the Michipicoten supracrustal rocks and intrusions of the Wawa domal gneiss terrane will be examined, followed by a look at the boundary between the Wawa terrane and Kapuskasing structural zone. Day three will be spent mainly in the Kapuskasing zone examining the Shawmere anorthosite complex, and high-grade gneisses, as well as the Ivanhoe Lake cataclastic zone separating rocks of the Kapuskasing zone from those of the Abitibi belt. On day four the geology of the Abitibi belt in the Timmins area will be outlined.

## Regional Setting

The Superior Province is an Archean terrane composed of east-west trending belts of alternate volcanic-rich and sediment-rich character, termed subprovinces (Fig. 1). The continuity of the east-west belts is interrupted by a northeast-trending zone of high-grade metamorphic rocks, the Kapuskasing structural zone (Thurston et al., 1977). The Kapuskasing structure is fault-bounded on the southeast but the western contact is complex and gradational over 120 km to low-grade rocks of the Michipicoten belt near Lake Superior (Figs. 1 and 2).

The Kapuskasing "high", a prominent northeasterly gravity and aeromagnetic anomaly, was interpreted by Wilson and Brisbin (1965) to indicate pronounced upwarp of the Conrad discontinuity. Bennett et al. (1967) concluded that the Kapuskasing structure is a complex horst uplifted during the Proterozoic. The association of 1,100-1,000 Ma

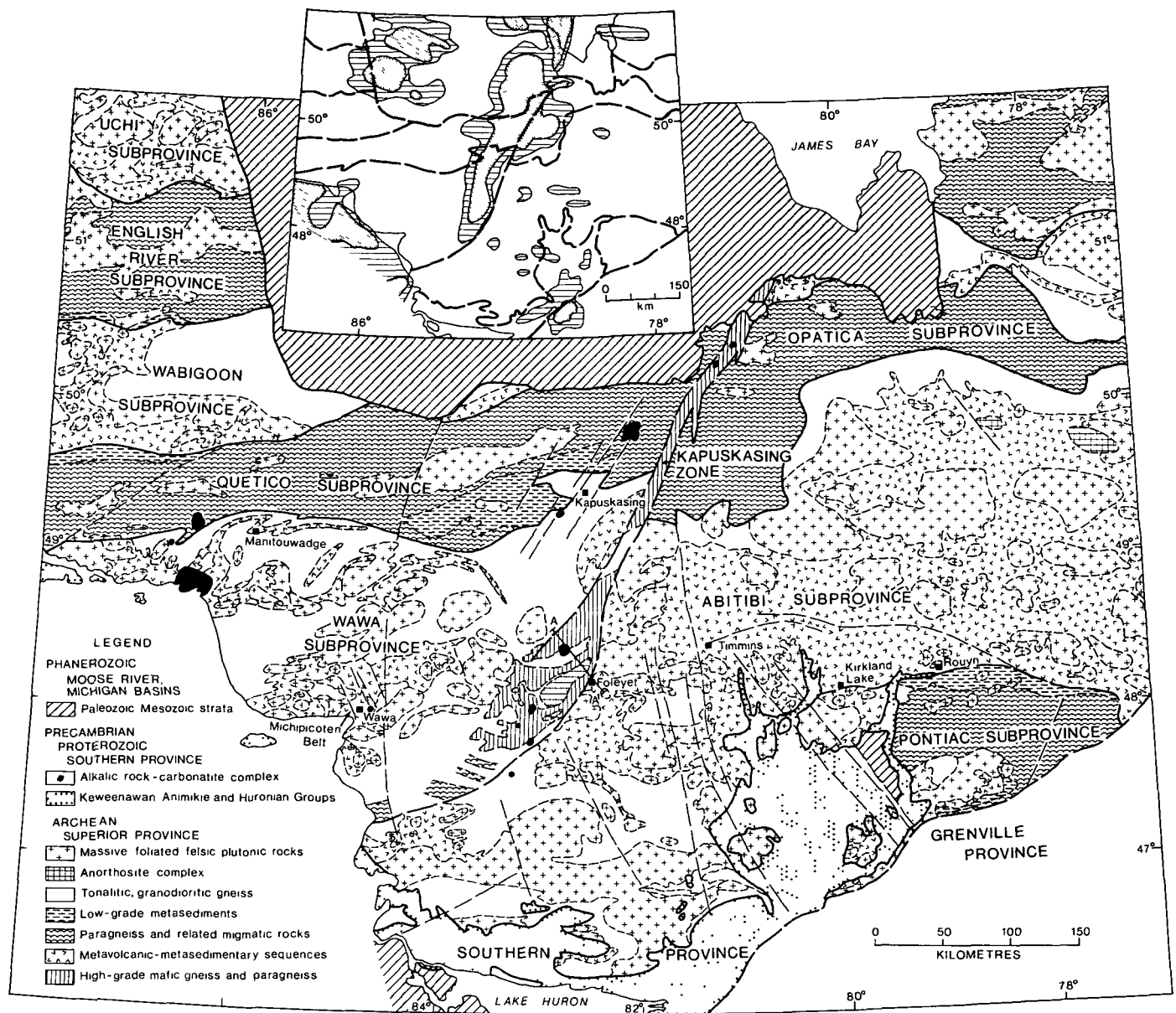
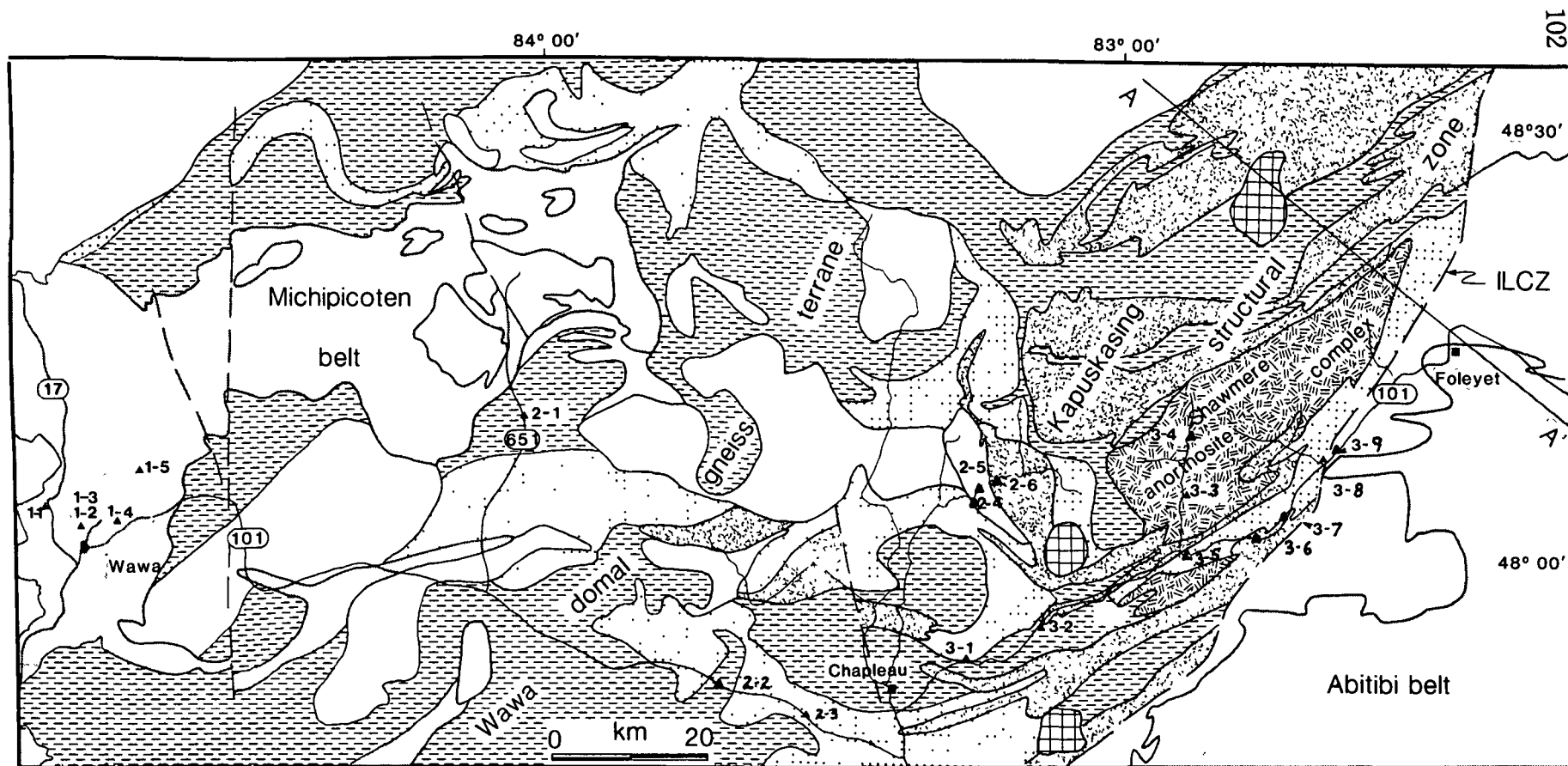


Fig. 1. Geology of the central Superior Province. Inset: positive gravity anomalies dark stipple ( $>-25$  mGal); lines ( $-25$  mGal to  $-35$  mGal); blank ( $<-35$  mGal).

alkalic rock-carbonatite complexes led Burke and Dewey (1973) to suggest that the Kapuskasing structure is a failed arm of the Keweenawan rift structure. Watson (1980) postulated that the Kapuskasing zone was uplifted during late Archean or early



# LEGEND

## Proterozoic

alkalic rock-carbonatite complex

## Archean

massive to foliated granite to tonalite

gneissic tonalite-granodiorite

xenolithic tonalite gneiss

metavolcanic, minor metasedimentary rocks

anorthositic rocks

paragneiss, mafic gneiss

— fault

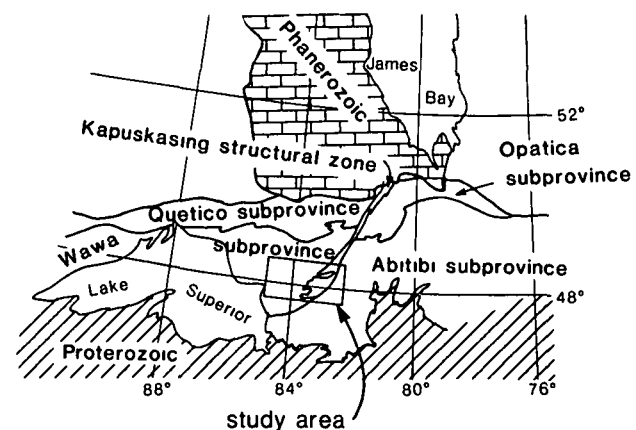


Fig. 2. Geology of the field trip route showing major rock units and tectonic subdivisions.

Proterozoic sinistral transcurrent movement. Recent earthquakes in the region indicate that the structure is still active (Forsyth and Morel, 1982).

#### Geophysical Characteristics of south-central Superior Province

A map showing apparent crustal thickness in the Lake Superior region, based on seismic refraction studies, was presented by Halls (1982) (Fig. 3). These data show that the apparent thickness of the crust of the Superior Province decreases easterly from values of about 45 km near Wawa to 35 km near Timmins. In addition, there appears to be a step-like decrease in depth to Moho associated with the eastern boundary of the Kapuskasing zone. Values in the 39 km range beneath the Kapuskasing zone drop abruptly to 35 km to the east. Halls viewed the contour map of apparent crustal thickness as preliminary, partly because of insufficient coverage in some areas and partly due to the necessity of averaging crustal and mantle velocities. Accounting for the high-density, presumably high-velocity rocks in the Kapuskasing zone would increase the crustal thickness estimates beneath the structure.

The Bouguer gravity anomaly map for the Wawa-Timmins region is shown in Figure 4. In general, areas underlain by metavolcanic rocks have associated positive gravity anomalies and granitoid-gneissic rocks have negative anomalies. The Kapuskasing structural zone has an associated linear positive gravity anomaly extending from James Bay in the north to some 50 km southwest of Chapleau. In the Wawa-Chapleau-Foleyet area, the gradient is gradual on the west and abrupt on the east, suggesting a west-dipping contact between the Kapuskasing zone and Abitibi subprovince. In this region, the gravity profile (Fig. 11) shows a paired high-low anomaly. The trough of the low is coincident with the fault at the eastern boundary of the Kapuskasing zone.

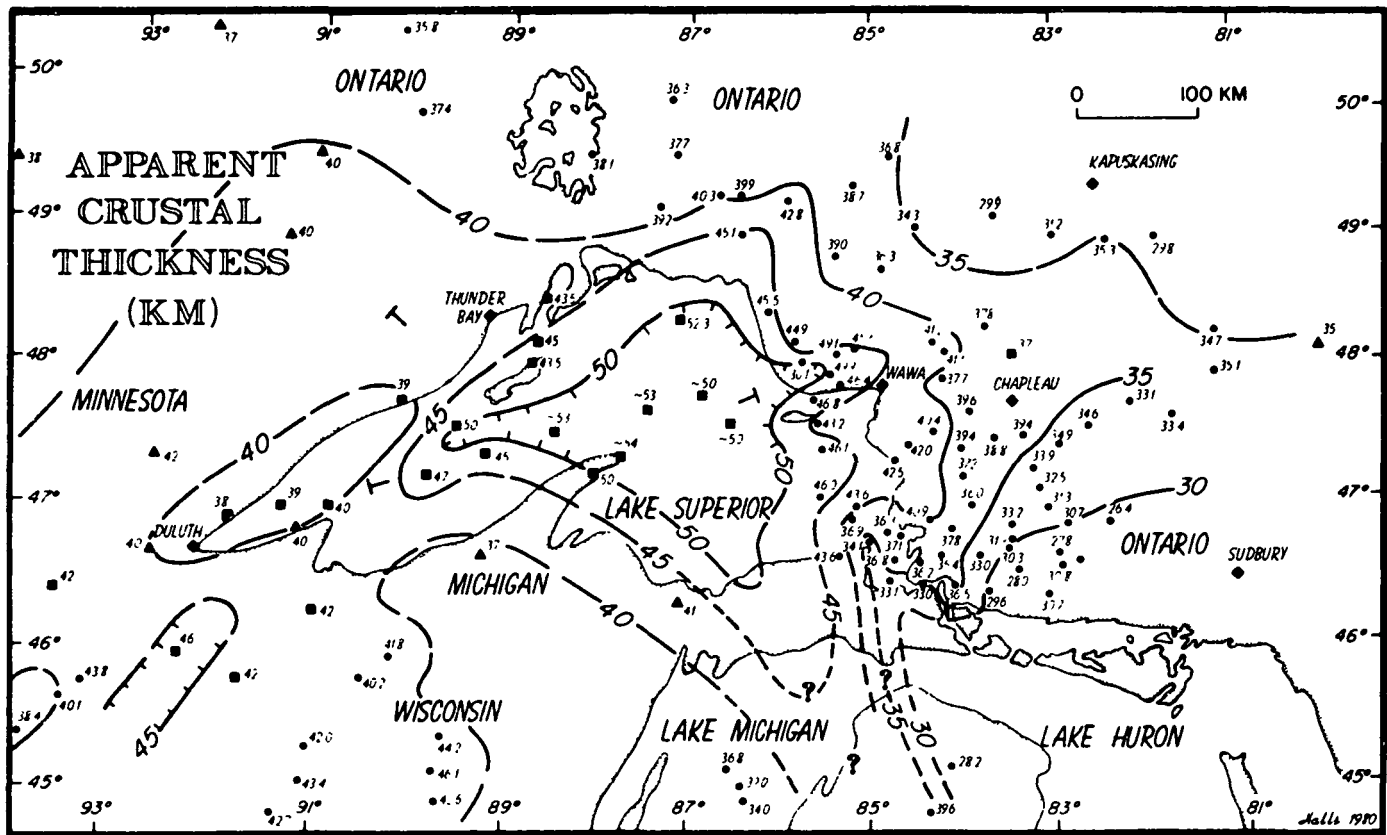


Fig. 3 Apparent thickness of the crust in the Lake Superior area (after Halls, 1982). Sources of data are listed by Halls and extreme caution in using map was advised.

To the north, the positive Kapuskasing anomaly broadens as it coalesces with the east-west gravity high associated with the Quetico-Opatika metasedimentary subprovince.

#### Geology of Wawa and Abitibi subprovinces

The geology of the region including the Michipicoten and Abitibi belts and Kapuskasing zone is shown in Figure 2. The Michipicoten belt, part of the volcanic-rich Wawa subprovince, is mainly composed of metavolcanic rocks of ultramafic, mafic and

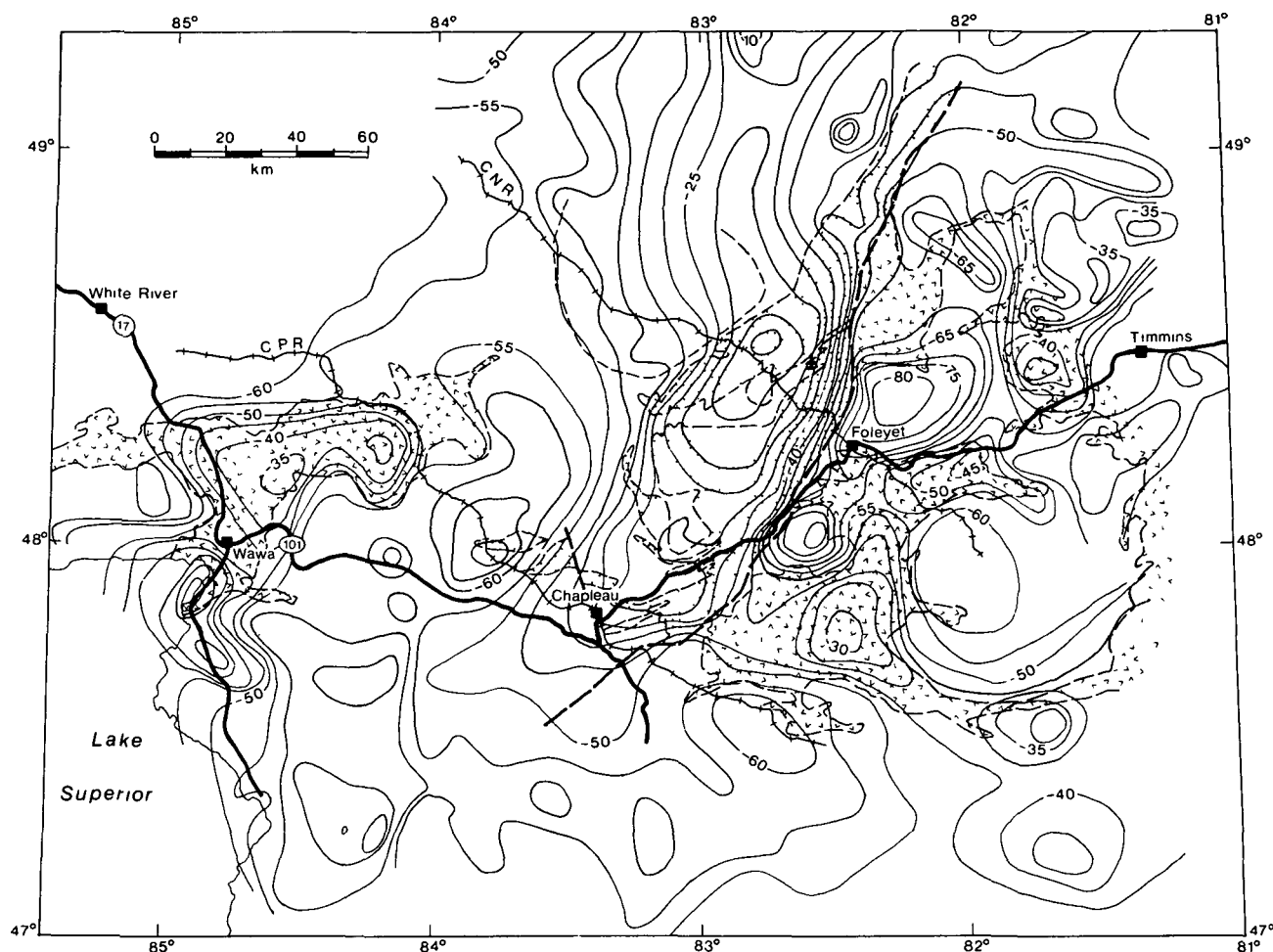


Fig. 4. Bouguer gravity map superimposed on simplified geology. Gravity from Earth Physics Branch map 44078, 44084, 48078 and 48084). Checks - greenstone belts; dots - Kapuskasing gneiss.

felsic composition (Goodwin, 1962), with intercalated greywacke, conglomerate, iron formation and chert. Dome and basin structures (Goodwin, 1962) as well as downward-facing strata and overturned structures (Attoh, 1980) have been recognized.

Metamorphic grade ranges from sub-greenschist to amphibolite facies

(Fraser et al., 1978). Several suites of intrusive rocks include synvolcanic bodies ranging from peridotite to granodiorite, younger granodiorite batholiths, and still younger granite and syenite plutons (Card, 1982).



The Michipicoten belt is intruded to the southeast by tonalitic gneiss and plutons of the Wawa domal gneiss terrane (Fig. 2). The rocks in this region consist of at least four lithologic components: (1) hornblende-plagioclase  $\pm$  clinopyroxene mafic and rare paragneiss xenoliths, ranging from centimetres to tens of metres in maximum dimension, making up 5 to 50% of individual outcrops, and enclosed in (2) the volumetrically most abundant phase, hornblende-biotite tonalitic gneiss which is cut by (3) concordant to discordant layers of foliated to gneissic biotite-hornblende granodiorite, which in turn are cut by (4) late discordant quartz monzonite pegmatite. Xenolith-rich tonalitic gneiss units alternate on a 5 to 10 km scale with xenolith-poor units and can be traced for distances of at least 50 km. Layering in mafic xenoliths is locally discordant to layering in enclosing gneiss. Small folds of layering in tonalitic gneiss are commonly truncated by layers of foliated granodiorite.

In the area between the Michipicoten belt and Kapuskasing zone (Fig. 2) the orientation of foliation, gneissosity and axial surfaces of small folds permit definition of several structural domains characterized by domal geometry. The spacing of major domal or antiformal culminations is on the order of 20 to 25 km, although many smaller culminations are also present. The Highbrush Lake and Racine Lake domes have cores of tonalite-granodiorite gneiss whereas the Chaplin Lake dome and Missinaibi Lake arch have granitic cores flanked by foliated to gneissic rocks. A planar fabric in the homogeneous granitic rocks, defined by lenticular quartz and biotite alignment, is generally concordant to gneissosity in mantling gneiss. The Robson Lake dome, adjacent to the Kapuskasing structural zone, has a core of interlayered mafic gneiss, paragneiss and tonalitic gneiss.

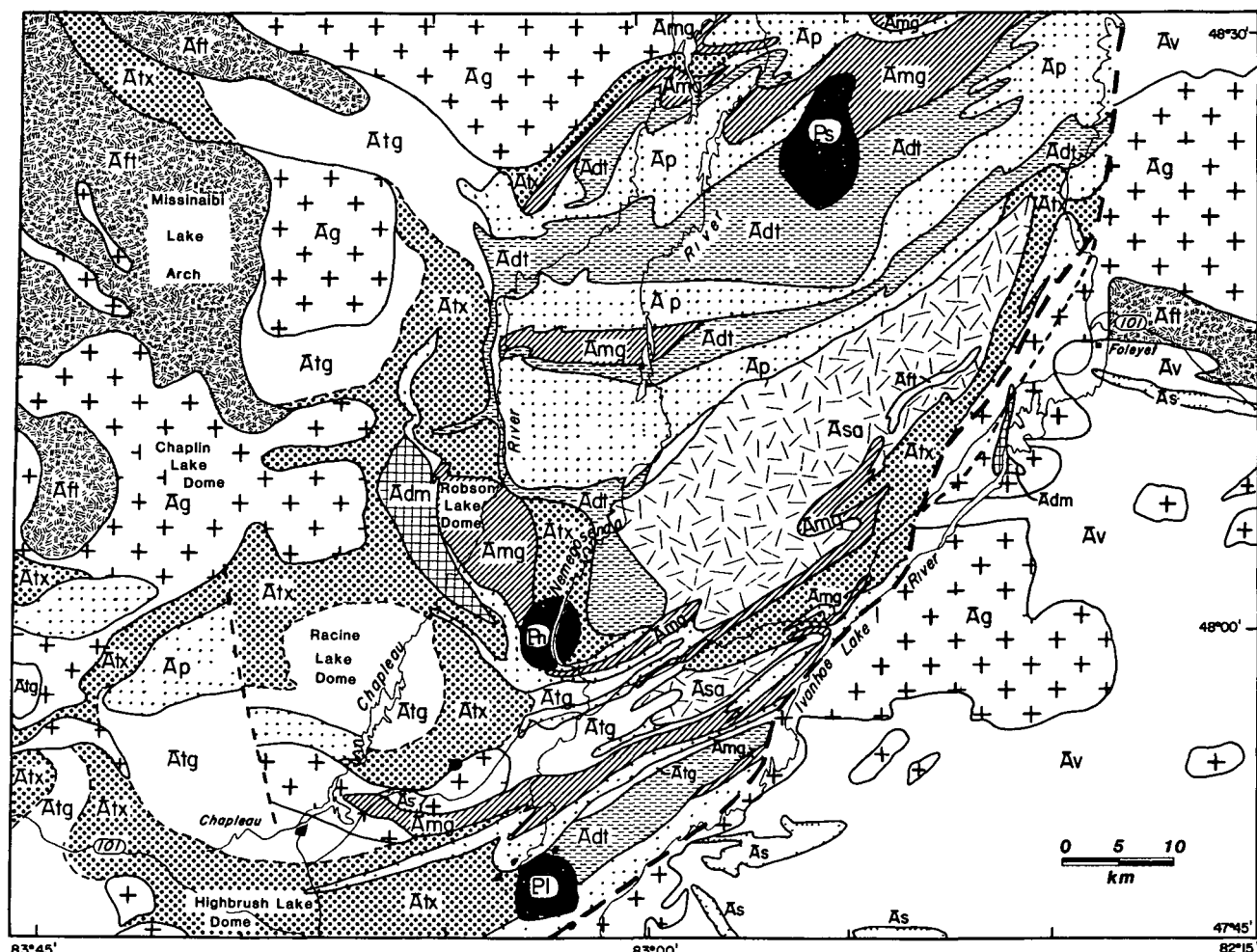
In general, asymmetric small folds of gneissic layering do not have a consistent sense of asymmetry with respect to domal culminations and are therefore not congruent with the domes. Near some domal crests, the orientation of gneissic layering, small

folds and lineations are widely variable to chaotic and define a pattern of coalescing domes.

Metasedimentary rocks occur in two locations in the eastern Wawa subprovince. A discontinuous, antiformal to domal belt of paragneiss west of the Racine Lake dome may be continuous to the east with paragneiss of the Kapuskasing zone (Figs. 2 and 5). Stretched-pebble metaconglomerate occurs in association with quartz wacke and amphibolite in the vicinity of Borden Lake. The polymictic (tonalite, granodiorite, meta-andesite, metasediments, amphibolite, vein quartz), clast-supported rock contains cobbles ranging from equant to constricted (1.5 m x 7 x 7 cm) with a prominent shallow northeast plunge. In cross-section the clasts vary from equidimensional to northwest-dipping ellipses.

The Floranna Lake complex is a strongly lineated and foliated complex crescentic pluton of intermediate composition that occurs between the Robson Lake and Racine Lake domes. The margins of the complex are fine- to medium-grained hypersthene-biotite granite, whereas the core contains medium-grained monzonite and diorite with rare gabbro and coarse biotite-clinopyroxene melagabbro layers. The least-deformed interior portions contain relict igneous(?) clinopyroxene and feldspar augen phenocrysts. Migmatitic quartz monzonite layers constitute up to 10% of some outcrops. The complex has similar structural and lithological characteristics to crescentic plutons of the Wabigoon subprovince of northwestern Ontario (Schwerdtner et al., 1979; Sutcliffe and Fawcett, 1979).

The eastern limit of the domal region is a semi-continuous zone of north, northeast and northwest striking, gently easterly-dipping gneissosity and easterly-plunging lineation. This curvilinear feature (Fig. 5) may represent the eastern extremity of a first-order dome of 75–100 km diameter, of which the individual structural domains



### Proterozoic



Alkalic rock-carbonatite complex: l: Lackner Lake complex;  
n: Nemegosenda Lake complex; s: Shenango complex

1100 Ma intrusions

### Archean



massive granite, granodiorite, with minor tonalite



diorite-monzonite intrusive complex, minor hornblendite, granite



foliated to flaser tonalite



tonalite-granodiorite gneiss; xenolithic



metavolcanic rocks, mainly metabasalt



metasedimentary rocks (includes metaconglomerate with tonalite cobbles with a U-Pb zircon date of 2664±12 Ma)



flaser diorite to mafic tonalite - includes minor gabbro, hornblendite, granodiorite



Shawmere anorthosite complex: metamorphosed gabbroic anorthosite, anorthosite, gabbro, minor tonalite



mafic gneiss: high Ca,Al basaltic composition, with tonalitic leucosome



paragneiss- quartz-rich composition, with up to 15% tonalitic leucosome

fault; Ivanhoe Lake cataclastic zone

2707-2668 Ma sequence

2749-2696 Ma sequence

pre-2765 Ma sequence

Fig. 5. Geology of the Kapuskasing structural zone and vicinity.

are higher-order domes of similar scale and spacing to those of the Wabigoon subprovince (Schwerdtner and Lumbers, 1980).

Dome development can be related in time to the formation of minor structures in gneiss. The discordant foliations in mafic gneiss predate the gneissic layering in the tonalite-granodiorite host. Small folds of this gneissic layering in turn predate intrusion of granodiorite layers. Crosscutting pegmatite dykes and sills are still younger and are probably the same age as the homogeneous plutonic rocks which locally have a planar fabric defined by lenticular quartz grains, biotite alignment, fracture cleavage, or minor planar zones of granulation. The absence of a consistent sense of asymmetry of small folds with respect to domal culminations and the random orientation of small folds near dome crests argue in favour of re-orientation of pre-existing small folds and gneissic layering during the latest doming. The quartz-lenticle foliation and fracture cleavage in homogeneous plutonic rocks cannot be readily attributed to magmatic flow and therefore suggest that the plutons were emplaced at their present structural level at sub-solidus temperature.

The Abitibi subprovince is dominated by a thick sequence of volcanic and sedimentary rocks of the Abitibi greenstone belt (Jensen, 1981). The supracrustal succession typically comprises sequences of ultramafic, mafic, and felsic volcanics. Intercalated turbiditic sedimentary rocks contain a high proportion of volcanic detritus. In the Abitibi belt, the uppermost group, the Timiskaming, is an unconformity-bounded sequence of alkalic volcanics and fluviatile sediments (Hyde, 1980) localized along major east-west fault zones.

Large areas of the Abitibi greenstone belt are metamorphosed to greenschist facies; subgreenschist, prehnite-pumpellyite facies rocks are common in the Timmins-Rouyn area and narrow aureoles of amphibolite facies rocks occur adjacent to plutonic bodies (Jolly, 1978).

The supracrustal rocks of the Abitibi subprovince display evidence of polyphase deformation in the form of major and minor structures of several ages and orientations. In the Abitibi greenstone belt, older northerly-trending folds are overprinted by east-west trending major and minor folds, forming major dome and basin structures (Pyke, 1982). The major isoclinal folds with east-west striking subvertical axial planes, steeply-plunging minor folds, subvertical axial plane foliation, and steeply plunging stretching lineation were probably formed under subhorizontal, generally north-south major compression. Toward the southern margin of Abitibi belt the major folds are overturned northward, and in the adjacent Pontiac subprovince, folds are recumbent. The Cadillac-Larder Lake fault zone, which constitutes the boundary between the Abitibi and Pontiac subprovinces, dips  $45^{\circ}\text{N}$  to  $60^{\circ}\text{N}$  and probably has both sinistral transcurrent and thrust components of movement.

Several suites of intrusive rocks in the Abitibi subprovince can be distinguished on the basis of composition, structural relationships, setting, and age (Card, 1982). The oldest suite includes synvolcanic sills, dykes and plutons ranging in composition from peridotite to granodiorite; the more felsic intrusions are typically quartz diorite and trondhjemite. Gneissic plutonic rocks of tonalite and granodiorite composition, commonly containing amphibolitic enclaves, occur in the northeastern and southwestern Abitibi subprovince. Massive felsic plutonic rocks intrude both the greenstones and the gneissic rocks in the form of simple and composite plutons and batholiths. They form several suites, including early granodiorites, younger granite batholiths, and still younger syenite-diorite plutons. Contacts between the plutons and the country rocks are commonly concordant and steeply dipping; dominant east-west structural trends are locally deflected about the intrusions.

A time framework for events in the Michipicoten and Abitibi belts can be constructed from U-Pb zircon dates. In the western Abitibi belt, volcanic rocks range in

age from 2,725 to 2,703 Ma (Nunes and Pyke, 1980; Nunes and Jensen, 1980) and in the Michipicoten belt, from 2,749 to 2,696 Ma, with synvolcanic plutons at 2,744 and 2,737 Ma (Turek et al., 1982). A number of late- to post-tectonic plutons from the Abitibi and Michipicoten belts have zircon dates within a few million years of 2,680 (Krogh et al., 1982). In the Wawa domal terrane, tonalite gneiss has a minimum age of 2,707 Ma, partly reset by intrusion of granodiorite at 2,677 Ma (Percival and Krogh, 1983; Fig. 5). Thus the Abitibi and Michipicoten supracrustal sequences and early intrusions developed between 2,750 and 2,700 Ma ago. The dates on volcanics and late plutons bracket the age of deformation and regional metamorphism at between 2,700 and 2,680 Ma ago. Major volcanic, plutonic, and tectonic events of relatively brief duration were essentially synchronous throughout the Abitibi and Wawa subprovinces, a region some 1,200 km long and 200 km wide. The lithologic and age similarities between the Abitibi and Wawa subprovinces strongly suggest original continuity, now interrupted by the Kapuskasing structural zone.

Diabase dyke swarms of late Archean and Proterozoic age are present throughout the region. The oldest dykes, the north-trending Matachewan swarm of the Abitibi subprovince, have a Rb-Sr age of 2633 Ma (Gates and Hurley, 1973). Northwest-striking diabase dykes in Wawa subprovince are petrographically similar to and have been paleomagnetically correlated with the Matachewan swarm (Ernst, 1981). Abitibi and Wawa subprovinces are thus inferred to have been tectonically stable cratons by this time. Northeast-striking tholeiitic dykes are about 2105 Ma old (Gates and Hurley, 1973); northwest olivine diabase dykes (Sudbury swarm) are about 1250 Ma old (Van Schmus, 1975); and east-northeast olivine diabase dykes (Abitibi swarm) are approximately 1100 Ma old (Lowden and Wanless, 1963).

### Kapuskasing Structural Zone

The Kapuskasing structural zone comprises northeast-striking, northwest-dipping belts of paragneiss, mafic gneiss, gneissic and xenolithic tonalite, and rocks of the Shawmere anorthosite complex (Bennett et al., 1967; Thurston et al., 1977) (Figs. 2 and 5).

Migmatitic paragneiss is compositionally layered with garnet, biotite, quartz-rich and rare graphitic varieties. Concordant tonalitic leucosome constitutes up to 20 per cent of many outcrops. Enclaves and layers of mafic gneiss in paragneiss occur on the 10 cm to 1 km scale. Migmatitic mafic gneiss is characterized by garnet-clinopyroxene-hornblende-plagioclase-quartz-ilmenite+orthopyroxene mineral assemblages and generally contains concordant tonalitic leucosome. Layering, on the 1 to 10 cm scale, is produced by variable proportions of minerals. Table 1 presents two sets of whole-rock analyses from adjacent anhydrous (garnet-clinopyroxene-plagioclase-quartz) and hornblende-bearing layers from mafic gneiss in two different locations. From the analyses it is unclear whether the layering is a preserved compositional heterogeneity or a product of metamorphic differentiation. The bulk composition corresponds to high calcium (10-15 wt% CaO), high alumina (13.4-17.2 wt%  $\text{Al}_2\text{O}_3$ ) basalt (Table 1). Nickel and chromium abundances of mafic gneiss are in the 95-220 and 12-190 ppm ranges respectively and are not definitive in distinguishing between basaltic igneous and marly sedimentary parentage for the rock type.

In the area of Figure 5, four linear, northeast-striking bodies of flaser-textured to foliated diorite and mafic tonalite occur dominantly within paragneiss terranes. These medium- to coarse-grained, locally migmatitic rocks consist of hornblende, biotite and plagioclase, with up to 10 per cent quartz as well as orthopyroxene, clinopyroxene and rare garnet. Gabbro, hornblendite and rare pyroxenite occur locally as layers 10 cm to 2 m thick, generally within 2 km of paragneiss contacts.

	1	2	3	4	5	6
SiO <sub>2</sub>	47.8	46.6	52.5	43.1	47.8	49.9
TiO <sub>2</sub>	0.81	0.81	1.81	1.59	1.0	1.3
Al <sub>2</sub> O <sub>3</sub>	15.5	15.6	17.2	13.4	16.2	17.0
Fe <sub>2</sub> O <sub>3</sub>	1.3	2.2	2.2	5.7	3.4	1.5
FeO	9.1	9.4	8.5	12.8	8.5	7.6
MnO	0.27	0.19	0.32	0.3	0.32	0.2
MgO	4.53	5.29	3.64	9.25	5.41	8.2
CaO	15.4	14.2	11.2	10.0	13.50	11.4
Na <sub>2</sub> O	2.0	2.4	2.8	1.6	2.3	2.8
K <sub>2</sub> O	0.25	0.41	0.12	0.58	0.33	0.2
H <sub>2</sub> O	0.5	1.1	0.3	1.6	0.8	
CO <sub>2</sub>	2.3	2.0	0.4	0.1	0.6	
Ni	0.014	0.014	0.0095	0.0098	0.024	
Cr	0.019	0.018	0.018	0.014	0.015	
Total	100.0	100.4	100.6	100.4	100.2	100.1

## CIPW Norm

QZ	1.5		6.6			
OR	1.49	2.44	0.71	3.47	1.95	1.0
AB	17.02	20.46	23.63	13.72	19.8	23.5
AN	32.77	30.81	33.92	28.03	32.97	33.4
DI	11.42	11.52	6.99	10.20	13.77	
HE	12.80	10.90	8.28	6.41	10.58	18.9
EN	6.05	3.64	5.80	5.39	4.96	
FS	7.83	3.95	7.89	3.89	4.43	9.4
FO		3.01		9.26	1.54	
FA		3.60		7.36	1.21	9.3
MT	1.9	3.21	3.18	8.37	4.95	2.2
IL	1.55	1.55	2.24	3.06	1.90	2.5
AP	0.12	0.14	0.26	0.24	0.19	
CC	5.26	4.58	0.91	0.46	1.44	

Table 1: Whole rock chemical analyses of mafic gneiss from the Kapuskasing zone, with CIPW norms. Analyst: R. Charbonneau, GSC Lab. 1: granulite layer, P79-475 (Gt-Cpx-Pl-Qz, 5% Hb); 2: amphibolite layer, P-475 (Gt-Cpx-Pl-Qz, 25% Hb); 3: granulite layer, P79-371 (Gt-Cpx-Pl, tr Qz); 4: amphibolite layer, P79-371 (Hb 40%, Gt 15%, Cpx 15%, Pl 20%); 5: average of three mafic gneisses from the KSZ (79-84A, 123, 299); 6: high-alumina basalt (Ringwood, 1975).



Discrete belts of xenolithic and gneissic tonalite are present south of the main body of the Shawmere anorthosite complex and small bodies are present to the north. The southern belt is made up of coarse garnet-hornblende-biotite-plagioclase-quartz tonalite containing enclaves of mafic gneiss, paragneiss, hornblendite and garnet-orthopyroxene-hornblende-biotite rocks. Southwest along this belt, garnet decreases in abundance and the composition is granodioritic. Inclusions in this area are amphibolite, hornblendite, and cummingtonite-hornblende-biotite rocks.

The Shawmere anorthosite complex (Thurston et al., 1977) consists of a main northern body, 15 x 50 km and a smaller mass, measuring 5 x 15 km. The bodies taper to the northeast and southwest and thus have concordant contacts. Gneissic textures prevail in the outer portions of the main body, whereas primary igneous minerals and textures are preserved in the interior (Simmons et al., 1980). The main body comprises four distinct lithological-textural units (Riccio, 1981; Fig. 6): (1) a border zone of migmatitic, foliated to gneissic garnetiferous amphibolite, (2) a banded zone consisting of 1 to 30 cm-thick layers of anorthosite, gabbro, garnet-rich, and ultramafic rock, (3) an anorthosite zone containing minor gabbro and (4) a megacrystic gabbroic anorthosite zone with plagioclase phenocrysts to 50 cm and minor anorthosite, anorthositic gabbro, gabbro and melagabbro. A 1 km wide body of foliated garnetiferous tonalite is present within the outcrop area of the anorthosite. Its genetic relationship to the anorthosite complex is not clear although it appears to be temporally related (Simmons et al., 1980). The southern body consists dominantly of coarse gabbroic anorthosite.

The orientation of gneissosity and lithological contacts make up the prominent east-northeast structural grain of the Kapuskasing structural zone. Gneissosity in all rock types is folded or warped about gently-plunging (0-25°) northeast-trending axes. The folds vary from isoclinal with consistent "Z" sense asymmetry when viewed toward the east to northwest-facing monoclinial flexures. Axial surfaces are rarely accompanied

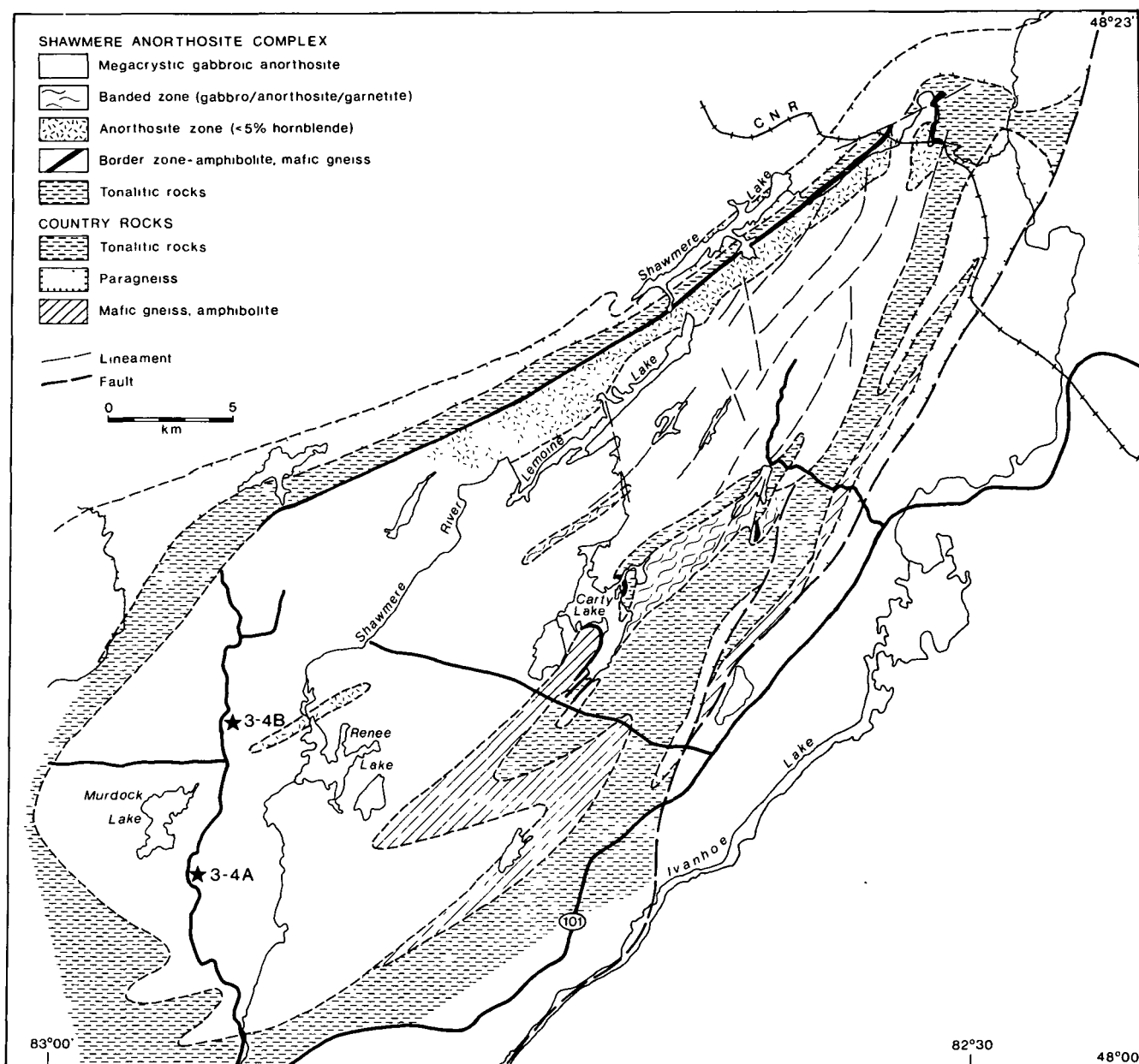
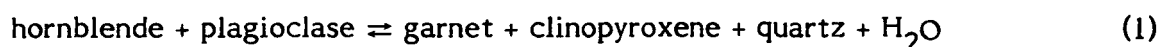


Fig. 6. Geology of the Shawmere anorthosite complex (after Riccio, 1980 and Percival, 1981).

by a foliation defined by flattened quartz grains. The trend of lineations and fold axes is northeast-southwest throughout this part of the Kapuskasing zone, but plunge direction varies on a regional scale from dominantly southeasterly in the south to northeasterly in the north. Between these areas, lineations are within 10° of horizontal and abrupt changes in plunge direction occur on the 10 m scale. Both regional and local plunge reversals can be related to a gently southeast-plunging warp axis.

Two high-grade metamorphic zones can be distinguished in this part of the Kapuskasing structural zone. Assemblages characteristic of a lower-grade garnet-clinopyroxene-plagioclase zone are developed in mafic gneiss. Orthopyroxene, present in four areas in most rock types, is diagnostic of a higher-grade orthopyroxene zone (Fig. 7; Percival, 1983).

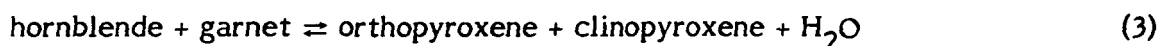
A continuous reaction resulting in decomposition of hornblende in mafic rocks to produce garnet and clinopyroxene may be written:



The coexistence over large areas of this divariant assemblage and tonalitic leucosome veinlets suggests that the reaction was anatectic and also produced a liquid over a range of P-T conditions (Fig. 8):

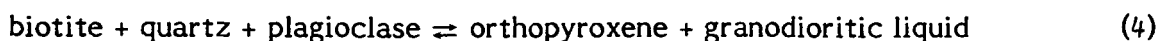


A possible reaction leading to the production of orthopyroxene in mafic rocks is:



The evolved water would presumably have been taken up by anatectic liquids.

In paragneiss, a reaction producing orthopyroxene in the presence of anatectic melt is:



A P-T diagram summarizing continuous reactions in the mafic system and apparent metamorphic conditions based on various mineral geothermometers and

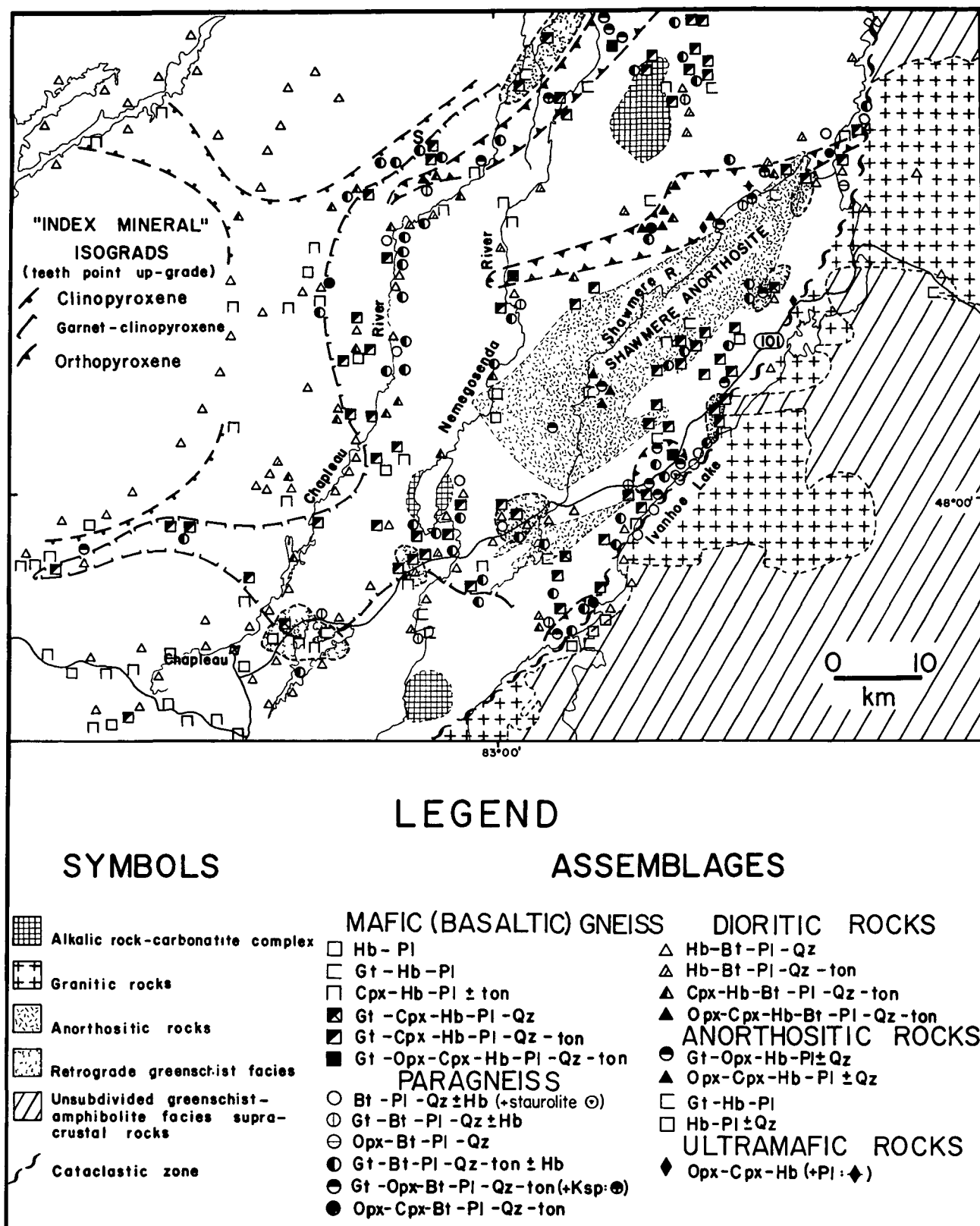


Fig. 7. Metamorphic mineral assemblages and index mineral isograds for part of the Chapleau-Foley area. Gt - garnet; Opx - orthopyroxene; Cpx - clinopyroxene; Hb - hornblende; Bt - biotite; Pl - plagioclase; Ksp - feldspar; Qz - quartz; ton - tonalitic segregations.

geobarometers, is presented in Figure 8. Apparent pressures, based on Newton and Perkins' (1982) garnet-clinopyroxene-plagioclase quartz barometer, are plotted on a map in Figure 9 and have an average value of 6.3 kbar. Apparent temperatures, based on the Ellis and Green (1979) garnet-clinopyroxene thermometer (Fig. 8) are in the range 700–800°C.

The assemblage almandine garnet-clinopyroxene-plagioclase-quartz is diagnostic of the regional hypersthene zone according to Winkler (1979, p. 260, 267–268). de Waard (1965) and Green and Ringwood (1967) suggested that this assemblage forms as an alternative to orthopyroxene-plagioclase during high-pressure granulite-facies metamorphism. Turner (1981) attaches a different significance to the assemblage, regarding it as transitional from amphibolite to granulite facies based on Binns' (1964) study. In the present study area, the location of the garnet-clinopyroxene-plagioclase zone between hornblende-plagioclase+clinopyroxene rocks and orthopyroxene-bearing rocks suggests that it characterizes the amphibolite-granulite facies transition. Although the assemblage is the same as that in the Adirondacks (de Waard, 1965) and temperature conditions were similar (cf. Bohlen and Essene, 1977), the path of metamorphism was different. In the Grenville Province, the development of garnet-clinopyroxene assemblages has been attributed to isobaric cooling of orthopyroxene-plagioclase granulites (Martignole and Schrijver, 1971; Whitney, 1978) whereas in the Kapuskasing zone, garnet and clinopyroxene formed during prograde reactions.

Rounded zircons of probable metamorphic origin from Kapuskasing mafic gneiss gave a concordant date of 2,650 Ma and from a leucosome layer in paragneiss of 2,627 Ma (Percival and Krogh, 1983; Fig. 10). A minimum age of emplacement for foliated tonalite from the Shawmere complex is provided by zircons (2,765 Ma) but the U-Pb system has been strongly affected by the high-grade metamorphism (Percival and Krogh, 1983). The rocks intruded by the tonalite are thus older than dated volcanic rocks of the Abitibi and Michipicoten belts.

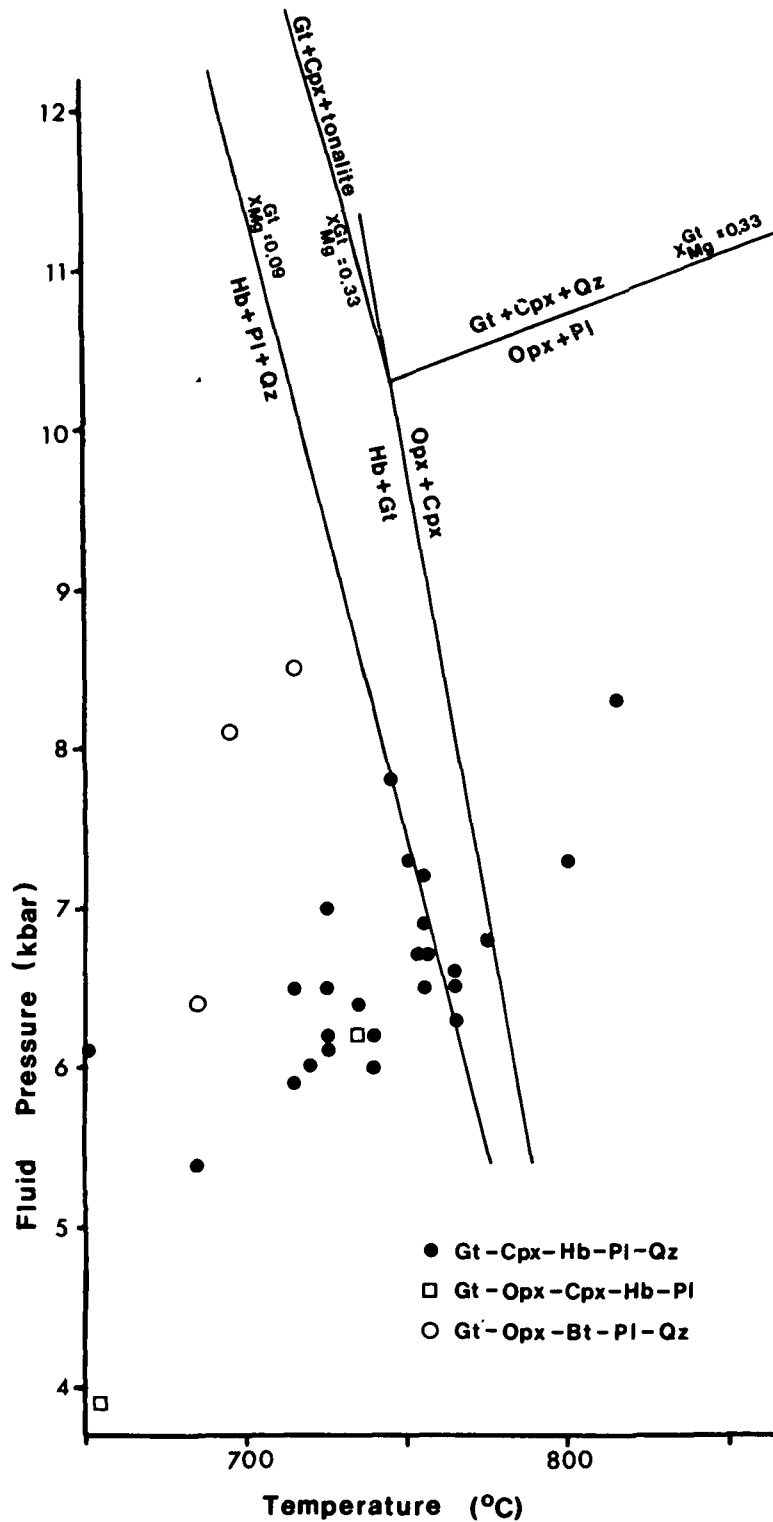


Fig. 8. Summary of reactions applicable to mafic rocks and metamorphic pressure-temperature estimates. Temperatures are derived from the garnet-clinopyroxene thermometer (Ellis and Green, 1979) and pressures from garnet-pyroxene-plagioclase-quartz barometers (Newton and Perkins, 1982).

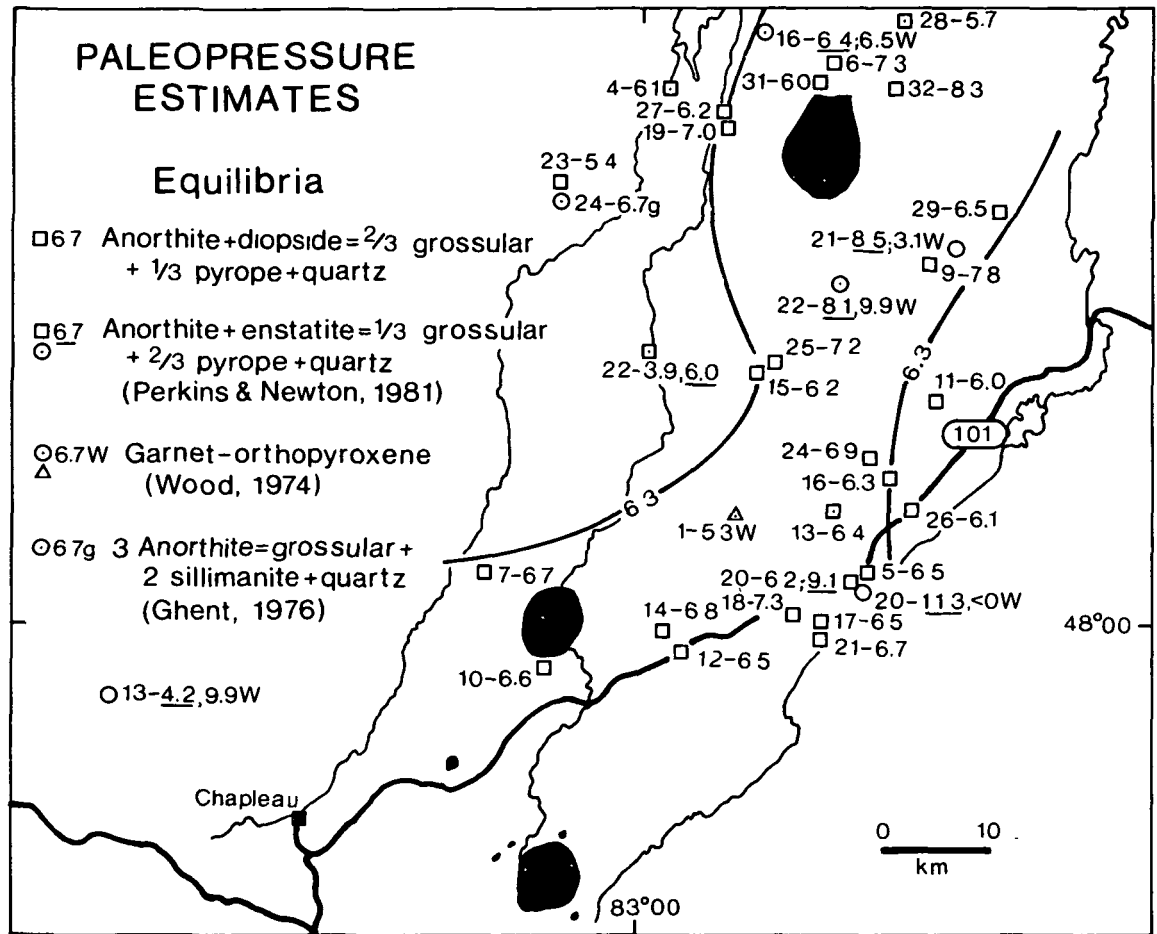


Fig. 9. Paleopressure map of the Chapleau-Foley area. Symbols represent rock type (circles - paragneiss; squares - mafic gneiss; triangles - orthogneiss). Numbers to the right of the dash are pressure estimators (kbar) keyed to the equilibrium used to derive the value. The 6.3 kbar reference line is based on garnet-clinopyroxene-plagioclase-quartz equilibrium.

At least two swarms of fresh mafic dykes transect metamorphic rocks of the Kapuskasing zone. East-northeast-striking, southeast-dipping Kapuskasing dykes are 1 to 10 m wide, sparsely plagioclase porphyritic, medium- to fine-grained, ophitic, green-grey gabbro. Northeast-trending olivine-bearing dykes may belong to the Abitibi swarm.





Several small alkalic rock-carbonatite complexes are associated with the Kapuskasing zone. The more northerly bodies have K-Ar dates of 1655 to 1720 Ma, whereas those in the south have dates of 1050 to 1100 Ma (Gittins et al., 1967). Thin lamprophyre dykes and a rare diatreme breccia are associated with the complexes; biotite from a lamprophyre dyke in the Chapleau-Foley area gave a K-Ar date of  $1144 \pm 31$  Ma (Stevens et al., 1982).

#### Relationship of Kapuskasing Structural Zone to Adjacent Subprovinces

The contact between the Kapuskasing structure and Abitibi subprovince is a zone of faulting and cataclasis, the Ivanhoe Lake cataclastic zone, that separates the two terranes of contrasting lithological, structural, and metamorphic characteristics. The contact is defined in part by a positive, linear north-northeast aeromagnetic anomaly and coincides with the trough of a paired high (Kapuskasing) - low (Abitibi) gravity anomaly (Figs. 4 and 11).

The Ivanhoe Lake cataclastic zone is characterized by narrow veinlets of finely comminuted rock which form discontinuous, randomly-oriented pods and networks. Two types of fault rocks can be distinguished. The first is foliated to massive, semi-opaque mylonite, cataclasite and blastomylonite, partly or totally recrystallized to fine grained epidote, chlorite, carbonate, and actinolite. The second type grades from cataclasite to pseudotachylite with aphanitic, almost opaque matrix and rounded, embayed monomineralic porphyroclasts.

The dip of the Ivanhoe Lake cataclastic zone is not well constrained geologically. Although some fault-rock veinlets are parallel to gneissosity and therefore dip gently northwest, many others have random orientation. The juxtaposition of high-grade against low-grade rocks indicates reverse displacement across the cataclastic zone. The associated paired gravity anomaly is characteristic of many well-documented overthrust terranes (Smithson et al., 1978; Fountain and Salisbury, 1981) and suggests that the

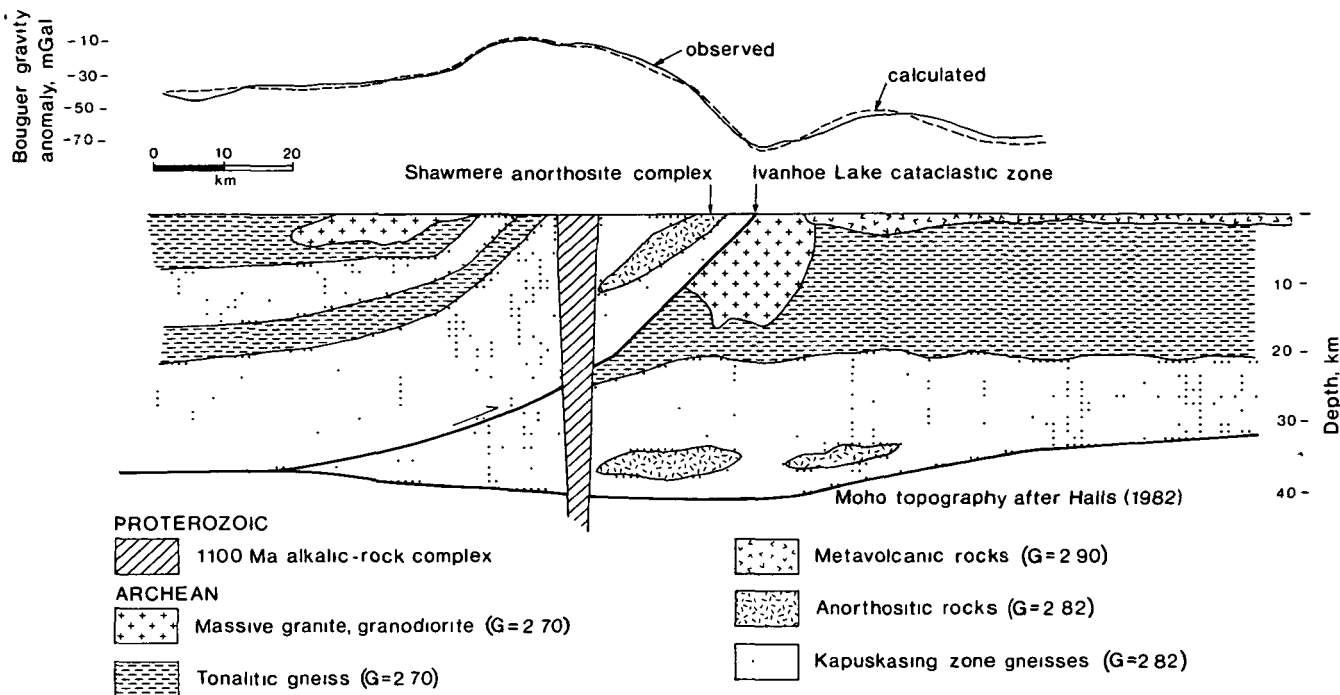


Fig. 11. Generalized west-east cross-section from the Wawa domal gneiss terrane, through the Kapuskasing structural zone into the Abitibi subprovince, showing gross crustal structure. The gravity model based on the average rock densities: tonalitic gneiss and granite: 2.70; metavolcanics: 2.90; Kapuskasing structural zone and lower crust: 2.82 g/cm<sup>3</sup>.

Ivanhoe Lake cataclastic zone is the surface expression of a northwest-dipping thrust fault (Fig. 11).

The Wawa-Kapuskasing boundary is gradational in lithological, structural and metamorphic characteristics. Mafic gneiss with minor paragneiss is typical of the Kapuskasing zone but also occurs in the Robson Lake dome with characteristic structural style of the Wawa subprovince. Garnet-clinopyroxene-hornblende-plagioclase assemblages are common here, suggesting that the metamorphic grade is similar to that in the Kapuskasing structural zone. The discontinuous paragneiss belt that extends for

up to 30 km into the Wawa subprovince may also be a part of the Kapuskasing lithological sequence. Tonalitic gneiss can be traced eastward from the Borden Lake area, where it has the complex structures characteristic of the Wawa subprovince, into strongly foliated and lineated gneiss typical of the Kapuskasing zone.

The change in structural style from domal in the eastern Wawa subprovince to linear ENE belts in the Kapuskasing structural zone can be used to define a transitional boundary zone, but no sharp line can be drawn on this basis. South of Chapleau, the orientation of gneissic layering changes eastward from horizontal near the Highbrush Lake dome, through a zone with a superimposed upright easterly foliation, to strong northeast-striking, northwest-dipping gneissosity. A north-south-trending structural culmination coincides with the eastern domes of the Wawa subprovince. East of the culmination, lineations plunge easterly toward a structural depression into which southwest-trending lineations of the southern Kapuskasing zone also plunge. To the north, lineations plunging northeasterly off the northeastern flank of the Missinaibi Lake arch appear to be continuous with northeast-plunging, reclined folds in the northern Kapuskasing structural zone. Cataclastic veinlets are sporadically present along mafic gneiss-tonalitic gneiss contacts for approximately 20 km southwest of Kapuskasing Lake. A fault zone marks the western limit of garnet-clinopyroxene-hornblende-plagioclase-quartz assemblages and mafic gneiss, but structural trends are continuous across it. The gradational nature of lithological contacts as well as the structural and metamorphic continuity between tonalites and high-grade gneisses suggests that the contacts were established prior to metamorphism and doming, and that rock units of the Kapuskasing zone locally occur structurally below the Wawa tonalite-granodiorite gneiss.

#### Structure of the Kapuskasing Crustal Cross-Section

The transition from the Michipicoten belt to the eastern boundary of the Kapuskasing zone can be interpreted as an oblique crustal cross-section based on the following: 1) metamorphic grade increases eastward from low greenschist facies in the

Michipicoten belt through amphibolite facies in the Wawa domal gneiss terrane to upper amphibolite and granulite facies in the Kapuskasing zone; 2) the proportion of plutonic to supracrustal rocks increases eastward in the Wawa subprovince; 3) the oldest rocks ( $> 2,765$  Ma) are in the Kapuskasing zone at the inferred base of the section; 4) the gravity anomaly can be best modelled by using a west-dipping crustal slab (Fig. 11).

Construction of a generalized crustal cross-section (Fig. 12) requires several assumptions: 1) the dip of the crustal slab is constant; 2) pressure is a function of depth so that estimates of metamorphic pressure can be used to derive the thickness of the section; 3) the metamorphic assemblages are the product of a single metamorphic event; and 4) post-metamorphic vertical displacement on faults within the section is negligible. The highest-grade assemblage from the Wawa area is garnet-andalusite in metagreywacke (Ayres, 1969), indicating a maximum pressure of 3.3 kb and a depth of about 11 km (Carmichael, 1978). The range of pressures estimated from the Kapuskasing zone, based on Newton and Perkins' (1982) garnet-clinopyroxene-plagioclase-quartz barometer, is 5.4 to 8.4 kb (average of 6.3 kb, Percival, 1983) but the lower values may result from re-equilibration during cooling. These values correspond to depths of 18 to 28 km (average 21 km). The minimum erosion-level difference is therefore 7 km, but the difference is probably closer to 15 km. The minimum and maximum dip estimates over a constantly-dipping slab 120 km long are approximately  $5^\circ$  and  $10^\circ$ .

The dips of post-metamorphic dykes in the Kapuskasing zone and eastern Wawa subprovince may provide an independent estimate of the tilt of the slab in this area. Matachewan dykes dip NE at  $75^\circ$  to  $85^\circ$  and ENE Kapuskasing dykes dip SE at  $70^\circ$  to  $85^\circ$  based on measurements of dykes with vertical exposure in roadcuts. Post-metamorphic mafic dykes in the Shield generally have near-vertical orientations, as do Matachewan dykes in the Abitibi subprovince (Thurston et al., 1977; Milne, 1972). The consistent non-vertical dip may thus have resulted from large-scale crustal rotation. To restore the dykes of both swarms to vertical, a  $14^\circ$  counter-clockwise rotation about an axis

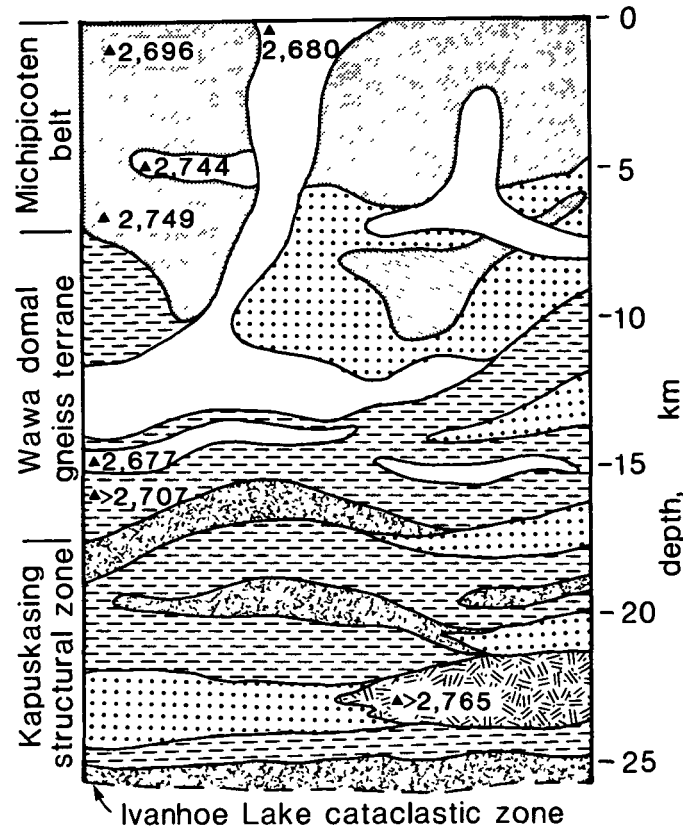


Fig. 12. Restored vertical section through the Michipicoten belt, domal gneiss terrane and Kapuskasing zone. Numbers are zircon dates ( $\pm 4$  Ma) on igneous and meta-igneous rocks.

trending  $038^\circ$  is necessary. Thus a  $14^\circ$  northwesterly dip is indicated in this eastern area. The difference in dip estimate provided by these two methods may be due to uncertainties in the data used in the calculations, faulty assumptions, or real differences in dip from east to west. The overall dip must flatten to the northwest and may in fact be reversed northwest of the Michipicoten belt. Ernst (1981, p. 87) reported consistent  $85^\circ$  SW dips of Matachewan dykes northwest of Wawa. Therefore, an intermediate dip value of  $10^\circ$  perpendicular to the fault was chosen for construction of Figure 12. If dips

flatten toward the northwest, this will result in over-estimation of the true thickness of the section.

The generalized section is a valid representation provided that (1) a single regional metamorphic event affected all of these rocks, and (2) late vertical displacement along faults is negligible between the Kapuskasing zone and western Wawa subprovince. In view of the complex relationships described and uncertainties involved, these simplifications may be unwarranted; however, the information which can be derived from an exposed cross-section through part of the crust is potentially valuable enough to permit some speculation.

The generalized crustal cross-section, constructed using a dip of  $10^\circ$  (Fig. 12), has at its base a sequence of upper amphibolite to granulite facies gneiss and anorthosite, the full thickness of which is unknown, and of which some 5 to 10 km is exposed in the Kapuskasing zone. Structurally above is 10 to 15 km of tabular batholiths of gneissic and xenolithic tonalite. Massive granitic rocks occur as sheets and deep-rooted plugs at this structural level. In the upper 5-10 km, both granitic rocks and gneissic migmatitic haloes surround the low-grade Michipicoten belt. The interfaces between the adjacent, generally horizontal megalayers are undulating surfaces with several kilometres of relief, manifest as gneiss domes at intermediate structural levels and as intrusive bodies at higher levels.

In the western Superior Province, two seismic discontinuities at 16-19 and 21-22 km, define upper, middle and lower crust (Hall and Brisbin, 1982). Using the Kapuskasing model, the upper discontinuity corresponds to the boundary between a structurally higher granitoid gneissic layer and a subjacent heterogeneous high-grade gneiss complex, whereas the lower discontinuity, corresponding to the middle-lower crustal boundary, is probably a metamorphic isograd (orthopyroxene isograd?) within the heterogeneous gneiss.

Similar models of mega-layered continental crust are based on seismic and gravity data (Smithson and Brown, 1977; Berry and Mair, 1980). Other inferred cross-sections through the crust (Ivrea zone, Pikwitonei region, Musgrave, Fraser ranges; Fountain and Salisbury, 1981) have in common a downward increasing metamorphic grade and a thick, intermediate-depth amphibolite-facies section of quartzofeldspathic gneiss, corresponding to the domal gneiss terrane of the Wawa subprovince. In the central Superior Province section, these gneisses intrude and assimilate both the overlying supracrustal succession and parts of the underlying complex. The entire section down to ~ 20 km was added to the crust in the interval between 2750 and 2680 Ma. The pre-existing crust may have, but need not have been as thick as present continental crust prior to the major thickening event. The high metamorphic grade in this older crust can be accounted for by burial, first by a volcanic pile and somewhat later by intrusion of tonalite sheets.

#### Archean Evolution of the Kapuskasing Crustal Structure

The oldest rocks so far recognized, paragneiss and mafic gneiss of the Kapuskasing zone, are considered part of a sedimentary-volcanic succession deposited prior to 2765 Ma ago. The Shawmere anorthosite was emplaced into this succession, probably also prior to 2765 Ma ago and probably as a stratiform body at depths of less than 20 km, as inferred from the presence of relict olivine (Thurston et al., 1977; Kushiro and Yoder, 1966). As suggested by Simmons et al. (1980), the intrusion may represent the differentiation product of tholeiitic basalt magmas which also erupted at surface.

Major eruption of volcanic rocks and deposition of sediments occurred between 2749 and 2696 Ma ago in the Michipicoten belt (Turek et al., 1982) and between 2725 and 2703 Ma ago in the western Abitibi belt (Nunes and Pyke, 1980). The lowermost volcanics are generally mafic and so have not been dated by the U-Pb zircon method.

Synvolcanic intrusions, including ultramafic, mafic, and trondhjemitic to granodioritic bodies, were intruded into the Michipicoten and Abitibi piles 2750 to

2700 Ma ago. Large volumes of tonalite intruded beneath and adjacent to the greenstone belts at this time. The minimum age of 2707 Ma for Wawa tonalite (Fig. 10) is given by a nearly concordant point and is therefore probably close to the true age. The tonalites could be the subsurface expression of magmas that produced dacites in the upper parts of the volcanic piles. Tonalite intrusions, now gneissic, engulfed and detached fragments of the lower parts of the greenstone succession (now represented as mafic xenolith trains), possible older, tonalite basement enclaves (e.g. Hillary and Ayres, 1980), and the western parts of the Kapuskasing zone which extend into the tonalite gneiss terrane. The tonalitic magmas may represent juvenile magmas derived from the mantle, or may be the products of partial melting of a heterogeneous lower crust similar to that exposed in the Kapuskasing zone. Heat from the tonalitic intrusions was probably sufficient to cause the metamorphism of the volcanics. Tonalitic magmatism thus may have coincided with regional metamorphism and acted as the main agent of heat transfer into the upper crust. Isoclinally folded gneissosity in the tonalite demonstrates that major deformation post-dates these intrusions.

The age of major deformation in the Abitibi and Wawa subprovinces is closely bracketed between 2696 Ma, the approximate age of the youngest volcanics, and 2680 Ma, the approximate age of late- to post-tectonic plutons (Krogh et al., 1982). In supracrustal rocks at high crustal levels, this deformation produced upright to vertically-plunging structural features as well as thrusts and nappe-like structures (Poulsen et al., 1981; Gorman et al., 1978; Thurston and Breaks, 1978). At deeper structural levels, the deformation resulted in gneissosity and subsequent folds in plutonic rock and paragneiss, followed by later doming. Forceful emplacement of massive plutons also deflected structural trends in country rock into concordance with the margins of these bodies. Following intrusion of the massive plutons at 2680 Ma, there was tectonic quiescence in Abitibi and Wawa subprovinces. There is evidence, however, of continued activity in the Kapuskasing zone.



High-grade metamorphic rocks of the Kapuskasing zone yield concordant U-Pb zircon dates of 2650 to 2627 Ma. U-Pb zircon dates are generally considered to record the age of crystallization of the zircons, which in this case are of metamorphic origin. This interpretation would imply that metamorphism in the Kapuskasing zone occurred 2650 to 2627 Ma ago, 25 to 50 Ma after tectonic stabilization of much of the rest of Superior province. A discrete burial and metamorphism event, restricted to the Kapuskasing zone, could explain the deformed metamorphosed conglomerate cobbles from Borden Lake which have a zircon date of 2664 Ma (Percival et al., 1981). However, tectonic mechanisms which could lead to deep burial of the 500 km long x < 50 km wide Kapuskasing "sliver" are unknown and seem to be unlikely after termination of the major tectonism in the Abitibi and Wawa subprovinces. This interpretation, that a second metamorphic event affected the area, is based entirely on geochronological results. Without these dates, a single metamorphic event would be adequate to explain the observed relationships.

One must therefore examine the assumption that zircons are closed to lead loss immediately following crystallization, regardless of the cooling history. Slowly decreasing metamorphic temperatures from peak levels of  $>800^{\circ}\text{C}$  could result in lead diffusion out of zircon for several million years after crystallization, provided that there is some finite "blocking temperature" for zircon. A value of  $700 \pm 50^{\circ}\text{C}$  was estimated for zircon blocking by Mattinson (1978). Invoking this hypothesis to explain the young "metamorphic" dates would allow a simpler geological history involving only the metamorphism at 2700 to 2680 Ma with slightly later ductile shear at depth.

The prominent east-northeast structural trends in the Kapuskasing zone are the result of relatively late tectonism. The structural grain is defined by the orientation of migmatitic and gneissic layering folded about shallow ENE axes. This folding event therefore post-dates crystallization of tonalitic melts, thought to coincide with the metamorphic peak. Similarly, structurally complex tonalitic gneiss units that can be

traced from Wawa subprovince into the Kapuskasing zone have a strong, superimposed ENE foliation and lineation in the Kapuskasing zone (Percival and Coe, 1981). If the correlation is valid between massive granodiorite dated at  $2680 \pm 3$  Ma and granodiorite gneiss adjacent to the Kapuskasing zone in the Abitibi subprovince, then the ductile strain occurred after 2680 Ma but pre-dated post-metamorphic cooling. This timing is consistent with the suggestion of Watson (1980) that sinistral transcurrent movement occurred along the Kapuskasing zone during emplacement of the Matachewan dyke swarm at 2633 Ma (Gates and Hurley, 1973). Late Archean deformation could have resulted in the resetting of 2700-2680 Ma zircons to ages in the range 2650-2627 Ma.

#### Uplift of the Kapuskasing Structure

The age of uplift of the Kapuskasing zone is not well constrained. Evidence of late Archean transcurrent movement was cited by Watson (1980) and Percival and Coe (1980), however its magnitude was probably small, judging by the minor apparent offset of the Abitibi-Opatika contact (Fig. 1). Major thrusting could also have occurred at that time, setting U-Pb and K-Ar isotopic systems in the high-grade rocks at 2,650-2,445 Ma.

Geochronological evidence indicates activity at 1,655-1,850 Ma. Three alkalic rock-carbonatite complexes near Kapuskasing have K-Ar dates of 1,655-1,720 Ma (Gittins et al., 1967). A biotite-whole-rock Rb-Sr isochron from tonalite of the Shawmere anorthosite complex is 1,850 Ma (Simmons and others, 1980). A whole-rock  $^{40}\text{Ar}/^{39}\text{Ar}$  analysis of blastomylonite from the Ivanhoe Lake cataclastic zone gave a date of 1,720 Ma (Percival, 1981; Fig. 13).

Three alkalic rock-carbonatite complexes in the southern Kapuskasing zone have K-Ar dates of 1,050-1,100 Ma (Gittins et al., 1967). Plagioclase from amphibolite in the footwall of the Ivanhoe Lake cataclastic zone yields a  $^{40}\text{Ar}/^{39}\text{Ar}$  plateau at 1107 Ma (Fig. 13), suggesting mild resetting, possibly due to faulting. Lower concordia intercepts of zircon discordia in the range 827-1,108 Ma (Percival and Krogh, 1983) may relate to uplift 1,100-1,000 Ma ago.

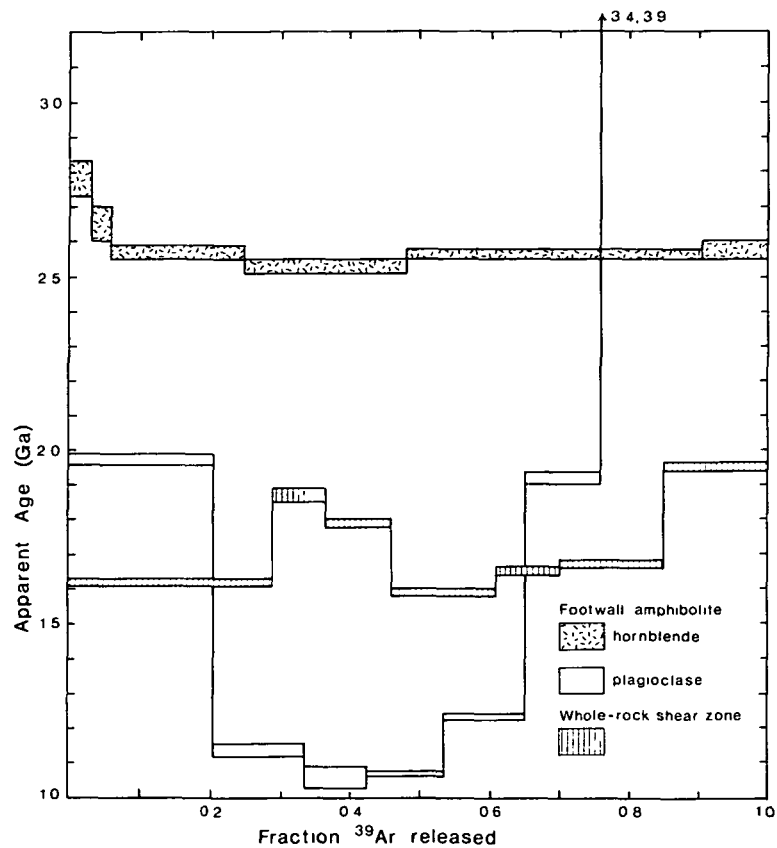


Fig. 13.  $^{40}\text{Ar}/^{39}\text{Ar}$  age spectra for hornblende, plagioclase and whole-rock separates from near the Ivanhoe Lake cataclastic zone. Increasingly high release temperature from left to right. Width of bars are  $2\sigma$  uncertainty estimates.

The coincidence of Proterozoic events along the Kapuskasing structure with major orogenic activity elsewhere in the Shield suggests that the structure is an intracratonic basement uplift related to a distant cause.

## DAY 1 - GEOLOGY OF THE MICHIPICOTEN GREENSTONE BELT, WAWA, ONTARIO.

The Michipicoten greenstone belt trends approximately east-west and consists of mafic and felsic flows and pyroclastics, clastic sediments, and iron formation, notably carbonate(siderite)-rich. U-Pb zircon ages on felsic volcanic units in the southern part of the belt indicate the sequence is about 2696 to 2749 Ma old (Turek et al., 1982). The supracrustal rocks are intruded and surrounded by felsic plutonic and gneissic rocks, including layered mafic-felsic tonalitic and granodioritic gneiss and granitic plutons of quartz diorite to granite and syenite composition. Several of the older, synvolcanic plutons yield U-Pb zircon ages of 2744 and 2737 m.y. A post-volcanic granodiorite is 2685 Ma old, and recently Turek et al. (1983) obtained an age of  $2888 \pm 2$  Ma on a granite southeast of the Michipicoten belt.

The rocks of the Michipicoten belt have been metamorphosed under low greenschist to amphibolite facies conditions, and have undergone extensive faulting and folding. There are major overturned (recumbent) fold structures with axial surfaces dipping 20° to 30° northeast (Attoh, 1980).

The supracrustal rocks of the southern part of the belt can be divided into three major cycles (Goodwin, 1962; Sage, 1980). A lower cycle, consisting of mafic and felsic volcanics, is capped by Michipicoten-type iron formation, mainly siderite, but with lesser pyrite-, chert- and graphite-rich rocks. The middle cycle comprises mafic volcanics overlain by clastic metasediments and felsic tuffs and breccias. The clastic sediments, including the Doré conglomerate, wacke, siltstone, and crossbedded arkose, are the facies equivalents of the felsic pyroclastics and are formed mainly of detritus eroded from the felsic centres. The upper cycle comprises intermediate to felsic (andesite - dacite) tuffs and quartz-feldspar porphyry. The associated Jubilee Stock, a high-level subvolcanic intrusion, was emplaced within a caldera structure (Sage, 1980).

## Michipicoten Belt Stop Descriptions

### STOP 1-1 - Doré Conglomerate

Exposures of Doré conglomerate will be examined on Highway 17, 9.5 km north of the junction of Highway 17 and 101 at Wawa.

The Doré is a thick, coarse polymictic conglomerate unit that is overlain and underlain by metavolcanic rocks. Eastward there is a facies transition from Doré conglomerate into a sequence of wacke, siltstone, cross-bedded arkose and conglomerate called the "Eleanor Slate". At this locality, coarse felsic tuff-breccias to the north are succeeded southward by a sequence of wacke (reworked tuff?) and Doré conglomerate. Bedding and foliation dip approximately  $45^{\circ}$  southeastward and may face downward (west).

The Doré consists of pebbles and boulders of mafic and felsic volcanics, quartz porphyry, iron formation and trondhjemite in a schistose, chloritic matrix. The pebbles are flattened in the foliation plane and elongated in the east-plunging rodding lineation. Variations in pebble packing and size define crude stratification units.

### STOP 1-2 - Helen Iron Range Section at McLeod Mine, Algoma Ore Properties Ltd.

The McLeod Mine extracts siderite and pyrite from the base of the Helen iron range. This iron range lies at the top of a 1800 m thick unit of intermediate to felsic metavolcanics consisting of oligomictic and polymictic breccia, thin bedded to massive tuffs, lapilli tuffs, spherulitic flows, flow banded flows, and crystal tuffs. These intermediate to felsic metavolcanics are intruded by gabbro to quartz diorite dykes and sills that reach 290 m in thickness.

The intermediate to felsic metavolcanics overlie a dominantly massive and pillowed sequence of intermediate to mafic metavolcanics.

The mafic to felsic metavolcanics are bimodal in composition and represent the oldest cycle of volcanism (Fig. 14).

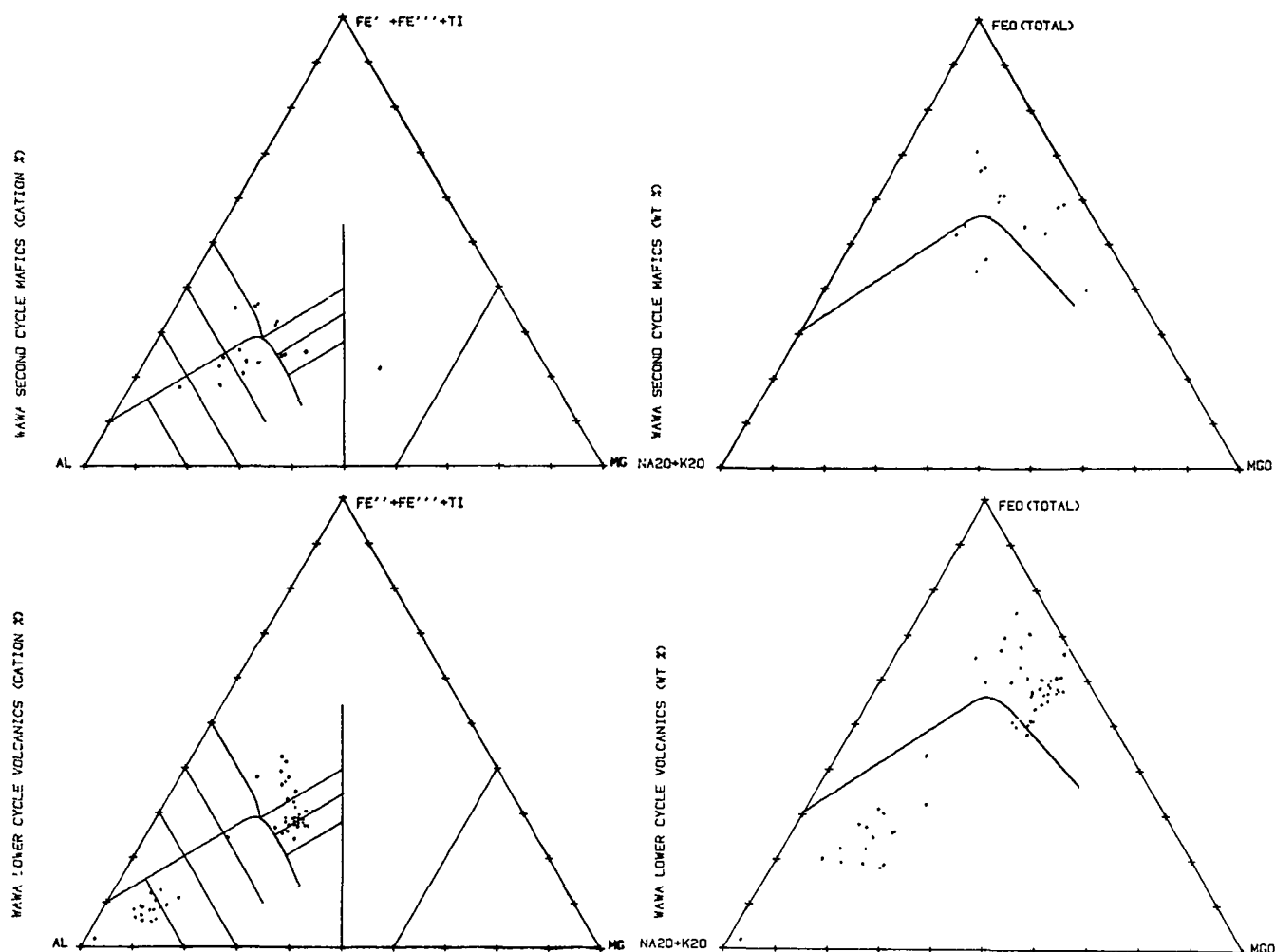


Fig. 14. AFM and Jensen plots for metavolcanic rocks of the Michipicoten belt.

The iron formation caps the first cycle of volcanism in the Wawa greenstone belt and from the stratigraphic bottom to top consists of five lithologic rock types that are gradational into each other. Upper and lower contracts are sharp. From stratigraphic bottom to top the iron formation consists of siderite, pyrite, banded chert, thin bedded chert-wacke-magnetite, and graphite-pyrite. A U-Pb isotopic age of approximately 2749 Ma has been obtained from the crystal tuffs lying immediately below the iron formation (Turek et al., 1982).

Overlying the iron formation are massive and pillowed intermediate to mafic metavolcanics of cycle two.

The area of the mine displays numerous major and minor faults, and contains several Proterozoic diabase dykes that strike northeast and northwest.

Beneath the iron formation an area of approximately 1800 m by 750 m has been outlined that contains visibly identifiable chloritoid. These crystals are generally up to 2 mm in diameter and are randomly oriented with respect to the schistosity. Chloritoid alteration occurs above the iron formation as well, but its presence is not visually as obvious.

The volcanics above and below the iron range are pervasively soaked with carbonate. Carbonate alteration is of regional extent and occurs in all rock types, volcanic, sedimentary, and intrusive.

### STOP 1-3

This stop contains the best exposure of a spherulitic (hollow) flow banded felsic flow within the McLeod Mine area. The spherulitic unit is overlain by a flow breccia containing good fiamme and this unit is in turn overlain by a massive tuff with scattered lapilli size clasts. This exposure lies in the lower part of the felsic part of the oldest cycle of volcanism. Stratigraphic tops are north and the section is overturned, dipping south. Interpreted faulting along Wawa Lake and beach deposits at Wawa prevent any estimate as to how far above the contact with the intermediate to mafic pillowed and massive metavolcanics this felsic section occurs.

### STOP 1-4a - Sir James Mine, Eleanor iron range

The Eleanor iron range is the stratigraphic equivalent of the Helen iron range and consequently the stratigraphic sequences of the two are essentially identical. The two ranges are separated by northwest-striking faults. In the Eleanor range, the intermediate to felsic metavolcanics of the stratigraphic footwall have been more

intensely sheared than the footwall rocks of the Helen iron range. As a result of this shearing, the footwall rocks of the Eleanor iron range are fissile and primary textures are not as well preserved. The exact thickness of the intermediate to felsic section is probably on the order of 510 m and has been intruded by dykes and sills of gabbro to quartz diorite.

The Eleanor iron formation is estimated to be approximately 75 m thick.

#### STOP 1-4b

The upper contact of the Eleanor iron formation consists of a unit of approximately 10 m thickness of graphite schist with pyrite nodules. This is the best exposure of the graphitic upper unit of the Michipicoten iron formation.

Lying in sharp contact with the stratigraphic top of the iron formation is the massive lower part of an overlying intermediate to mafic flow. The graphite unit is cut by an irregular quartz porphyry dyke.

Immediately west of the exposure of the upper contact of the iron formation excellent exposure of pillowed intermediate to mafic metavolcanics. These intermediate to mafic metavolcanics are the base of cycle two volcanics. Note that from pillow shape the sequence faces north.

#### STOP 1-5 (optional) - Lucy Iron Range

The Lucy iron range is the faulted eastern extension of the Eleanor iron range and is thus stratigraphically equivalent. Left lateral movement along the Midland Lake fault, which separates the two iron ranges, is approximately 3.1 km.

The stratigraphy of the Lucy range is identical to that of the Helen and Eleanor iron ranges and the intermediate to felsic part of the section is of a thickness similar to that of the Eleanor iron range.

The thickness of the underlying massive and pillowed intermediate to mafic metavolcanics of the first cycle is unknown due to faulting. At the Lucy range the mafic



and felsic parts are separated by a polymictic mafic breccia of approximately 50 m thickness. The overlying intermediate felsic metavolcanics consist of lapilli tuffs, bedded tuffs, and a rare fiamme-bearing flow. The intermediate to felsic section is approximately 250 m thick and has been intruded by dykes and sills of possibly original gabbro composition.

The felsic unit is overlain by iron formation of an estimated 75 m width. The stratigraphy of the iron formation is the same as that of the previous iron ranges. The upper section of the Lucy range contains good thin bedded chert-wacke. The wacke beds display grain gradation, and numerous interformational units of brecciated chert-wacke occur. The brecciated units alternate with unbrecciated units; the breccias are primary in origin.

Above the iron formation lie approximately 1000 m of monotonous intermediate to mafic massive and pillowed mafic volcanics of cycle two.

Cycle two mafic volcanics are overlain by metasediments that are dominantly wackes. The volcanic-sedimentary contact is conformable and defined by graphite-pyrite, and pebbly conglomerate with pyrite. Immediately overlying the graphite-pyrite and pebble conglomerate are thin bedded wackes that are presently chlorite schists. Above the chlorite-schist sediments lie wackes that approach arkose in composition. These cleaner wackes display excellent cross-bedding and rare ripple marks, and dessication cracks have been found. The thickness of the overlying metasediments is unknown due to faulting, but at least 450 m is present. The cleaner wackes may be the sedimentary detritus from the felsic centre of cycle two volcanism located 4 km northwest of the McLeod Mine. This centre has been dated by U-Pb methods at approximately 2696 Ma (Turek et al., 1982).

## DAY 2 - GEOLOGY OF THE WAWA DOMAL GNEISS TERRANE IN THE WAWA-CHAPLEAU AREA

In this section we will examine the relationships between supracrustal rocks of the Michipicoten belt and plutonic rocks of the Wawa terrane, and the transition from Wawa domal gneiss to high-grade rocks of the Kapuskasing zone.

km

00.0 Junction of Highway 101E and Mission Road, Wawa. Proceed east on Highway 101.

63.1 Junction Highway 101 and Highway 651. Proceed north on Highway 651.

### 80.2 STOP 2-1 - Mafic gneiss

A large enclave of mafic gneiss is enclosed in and intruded by tonalitic gneiss in a migmatitic zone marginal to the Michipicoten greenstone belt. The mafic gneiss is considered to represent deformed, metamorphosed (amphibolite facies) Michipicoten volcanics. The clinopyroxene-hornblende-plagioclase mafic gneiss is compositionally layered (1 to 40 cm) and contains thin (1 to 5 mm) concordant tonalite layers. It is cut by early tonalitic intrusions, late aplitic and pegmatitic dykes, and still later mafic and lamprophyric dykes. The gneiss displays subvertical foliation, mineral lineation, and tight steeply-plunging isoclinal minor folds.

Return to Highway 101 and proceed east.

### 127.1 STOP 2 - Tonalite gneiss and Mafic dykes

Tonalite gneiss is cut by northwest- and northeast-trending mafic dykes with good chilled margins. The older northwest-striking Hearst dykes (Ernst and Halls, 1980) occur west of the Kapuskasing zone. The dykes have a similar trend and similar characteristics, including plagioclase phenocrysts and tholeiitic composition, to Matachewan dykes east of the Kapuskasing zone. The Matachewan dykes have an Rb-Sr whole-rock age of  $2633 \pm 75$  Ma (Gates and Hurley, 1973). Ernst and Halls (1980) also reported similar paleomagnetic poles for the two swarms. In a zone 50 km wide west of the Kapuskasing zone, the Hearst dykes average 4 m in width and have a consistent easterly dip of  $80^\circ$  (Ernst, 1982; Percival, 1981). The tonalitic gneiss is thinly layered

and has sparse mafic xenoliths. The structure, although appearing chaotic, is generally subhorizontal. There is evidence for at least two foliations, and older gneissosity that is reoriented by a younger foliation to give complex sigmoidal patterns.

km

#### 153.6 STOP 2-3 - Highbrush Lake Dome

Small domes exposed in roadcuts here are small-scale examples of the regional-scale structural pattern. The main rock type is fine grained, thinly layered biotite and hornblende-biotite gneiss with local augen of potash feldspar.

The outcrop consists of several domes or canoe-shaped folds. Small intrafolial folds are present in some parts of the outcrop, as are amphibolitic xenoliths, locally with clinopyroxene. Pink granitic pegmatite dykes and sills cut gneissic layering in some areas.

#### 161.7 Junction of Highway 101 and Highway 129.

Proceed north on Highways 101-129 through Chapleau. Follow the Missinaibi Provincial Park signs north of town on gravel road. Proceed north and east on park road and logging roads leading to the Chapleau River.

#### STOPS 2-4 and 2-5 - Floranna Lake Complex

The Floranna Lake Complex is a complex crescentic pluton of intermediate composition. The western margin consists of lineated, fine grained granite with orthopyroxene and biotite. Inside the marginal unit is lineated diorite to monzonite containing hornblende-rimmed augen clinopyroxene phenocrysts. The central part of the complex is foliated and lineated hornblende-clinopyroxene-biotite diorite, gabbro and melagabbro. The well-exposed eastern contact of the complex shows extremely attenuated and contorted layering in granite of the complex adjacent to rocks of the Robson Lake dome to the east.

#### 208.4 STOP 2-4 - Clinopyroxene augen diorite-monzonite

These rocks are strongly lineated, medium- to coarse-grained monzonite. Rod-shaped clinopyroxene phenocrysts, rimmed by hornblende, make up some 20% of the

rock. Original outlines of feldspar grains are visible but these are now polycrystalline aggregates.

km

#### 209.4 STOP 2-5

Foliated and lineated diorite with gabbro and melagabbro layers. Igneous clinopyroxene with hornblende overgrowths and granitic leucosome layers are present.

#### 213.3 STOP 2-6 - Granulite gneiss, Robson Lake dome

The easternmost exposure of the igneous complex is strongly foliated, hornblende porphyritic granite augen gneiss with leucogranite veinlets parallel to foliation on the 1 mm-2 cm scale. The layering is folded about a gently northwesterly-plunging axis and the eastern limb is fine-grained and mylonitic with highly attenuated layering.

To the west, the Robson Lake dome consists of interlayered mafic and tonalitic gneiss. Near the contact with the Floranna Lake Complex, the layering in rocks in the dome is concordant to the contact and dips steeply west, however, the attitude is horizontal farther east in the core of the dome. At this outcrop the mafic rocks consist of garnet-clinopyroxene-hornblende-plagioclase-quartz assemblages, with layering defined by different proportions of minerals, including garnet-rich and hornblende-rich varieties. Concordant tonalitic layers locally have large clinopyroxene crystals rimmed by hornblende, in clots up to 4 cm. The layering is folded about upright isoclinal folds locally.

Small dykes of granite in mafic gneiss and inclusions of mafic gneiss in granite indicate an intrusive contact between the Floranna Lake Complex and Robson Lake dome.

The interpretation of the Robson Lake dome, which has the structural attributes of the Wawa domal gneiss terrane and the lithological characteristics of the Kapuskasing zone, is that rocks like those exposed in the Kapuskasing zone underlie the Wawa domal terrane and have been exposed here in a structural culmination.

### DAY 3 - KAPUSKASING STRUCTURAL ZONE

In this section we will examine the rocks and structures of the Wawa domal gneiss - Kapuskasing zone boundary and of the Kapuskasing zone itself along Highway 101.

km

0.0 Junction of Highway 101E and Highway 129 south of Chapleau. Proceed east on Highway 101E.

#### 13.0 STOP 3-1 - Borden Lake conglomerate

This outcrop consists of stretched-pebble metaconglomerate with a strong rodding lineation and weak, gently north-dipping foliation. The rock is a clast-supported conglomerate containing ~10% matrix of garnet-hornblende-biotite-quartz. The cobbles, which range up to 1 m in length, are felsic metavolcanics, metasediments, granodiorite-tonalite, plagioclase-porphyrritic meta-andesite and amphibolite, with rare hornblende and vein quartz. The metaconglomerate is spatially associated with amphibolite and paragneiss to the south on Borden Lake, and is cut by granite, however, the stratigraphic relations of the supracrustal rocks are unknown.

Tonalitic cobbles extracted from the metaconglomerate have yielded zircons dated at  $2664 \pm 12$  Ma (Percival et al., 1981). The zircons have a corroded appearance and produced discordant data points and hence the interpretation of the data is open. The zircons could preserve the original crystallization age of the source pluton for the cobbles or they could record a later deformation-metamorphic event.

#### 25.9 STOP 3-2 - Mafic gneiss xenoliths with amphibolitic retrograded margins

This exposure demonstrates an important aspect of the boundary between the Kapuskasing zone and Wawa gneiss terrane. Aside from the large-scale structural contrast between the domal Wawa terrane and the linear belts in the Kapuskasing zone, intrusive relations are also instructive. The outcrop consists of two main components: (1) coarse-grained hornblende-biotite tonalite, dominant to the west, and (2) medium

grained mafic gneiss consisting of garnet-clinopyroxene-hornblende-plagioclase-quartz assemblages. Small xenoliths of mafic gneiss in tonalite have margins, up to several cm thick, consisting of hornblende-plagioclase. Dykes of tonalite cutting mafic gneiss are bordered by mafic rock with hornblende-plagioclase assemblages. The interpretation of age relationships is that the high-grade metamorphism that produced the garnet-clinopyroxene assemblages in mafic gneiss preceded the intrusion of tonalite. Water in the tonalite margin was presumably released upon crystallization and hydrated the adjacent less-hydrous mafic rock. Although the tonalite at this outcrop has not been dated, it probably belongs either to the >2707 Ma tonalite gneiss suite or to the 2680 Ma group of plutons. The high-grade metamorphism is therefore older than 2680 Ma. This is in conflict with the direct dates of metamorphic zircons from the Kapuskasing zone that yield ages of  $2650 \pm 2$  and  $2627 \pm 3$  Ma. The conflict has led to the suggestion (Percival and Krogh, 1983) that the Kapuskasing gneisses, although metamorphosed prior to 2680 Ma ago, remained at high temperatures where radiogenic lead diffused readily out of zircon until <2627 Ma ago.

35.6 Paul Township road - turn left and proceed north.

#### STOPS 3-3 and 3-4 - Shawmere anorthosite complex

The Shawmere anorthosite complex is a layered calcic anorthosite body. In this traverse we will examine the central portion comprising mainly megacrystic gabbroic anorthosite.

#### 46.8 STOP 3-3

This outcrop consists of coarse grained gabbroic anorthosite with patches of anorthosite and gabbro on the 1 m scale. Original outlines of plagioclase ( $An_{50}$ ) phenocrysts, now polycrystalline aggregates, are up to 10 cm. Local mafic patches contain the assemblage plagioclase ( $An_{92}$ )-garnet-orthopyroxene-hornblende-gedrite-spinel-sapphirine (Table 2). Garnet is present both as discrete grains and in coronal structures between hornblende and plagioclase.

	OG-1 1	OG-1 2	OG-1 3	OG-1 4	OG-1 5	OG-1 6
SiO <sub>2</sub>	41.18	53.67	12.80	0.26	44.04	44.29
TiO <sub>2</sub>	0.03	0.00	0.08	0.03	0.10	0.05
Al <sub>2</sub> O <sub>3</sub>	23.01	4.66	62.27	64.11	19.04	16.02
Cr <sub>2</sub> O <sub>3</sub>	0.09	0.05	0.06	0.06	0.08	0.08
FeO	19.70	13.48	6.33	19.09	11.01	7.80
MnO	0.55	0.12	0.06	0.06	0.18	0.08
MgO	14.15	27.74	16.75	15.43	21.11	15.50
CaO	3.38	0.07	0.04	0.00	0.62	11.00
Na <sub>2</sub> O	0.44	0.65	0.00	0.07	2.03	1.74
K <sub>2</sub> O	0.01	0.08	0.00	0.01	0.00	0.21
Total	102.60	100.56	98.41	99.21	98.22	96.82
Si	2.981	1.894	0.769	0.007	6.115	6.273
Ti	0.002	0.00	0.004	0.001	0.010	0.005
Al(iv)	0.00	0.108	0.231	0.00	1.885	1.728
Al(vi)	1.962	0.088	4.176	1.974	1.230	0.946
Cr <sub>3</sub>	0.005	0.001	0.003	0.002	0.008	0.011
Fe <sub>2</sub>	0.00	0.065	0.034	0.002	0.065	0.474
Fe	1.192	0.333	0.284	0.415	1.214	0.450
Mn	0.034	0.004	0.003	0.001	0.021	0.010
Mg	1.526	1.459	1.499	0.601	4.368	3.271
Ca	0.262	0.003	0.003	0.00	0.092	1.699
Na	0.062	0.044	0.00	0.004	0.546	0.478
K	0.001	0.004	0.00	0.00	0.00	0.038

**Table 2** Microprobe analyses of minerals in magnesian melagabbro from Shawmere anorthosite complex. 1: garnet; 2: orthopyroxene; 3: sapphirine; 4: spinel; 5: orthoamphibole; 6: clinoamphibole. Also present is anorthite (An<sub>94</sub>).

km

#### 54.7 STOP 3-4

At this exposure, large crystals of orthopyroxene, partly rimmed by hornblende and garnet, are preserved in anorthositic gabbro. The more mafic layers alternate with plagioclase-rich layers on a 20 cm scale. The effect of deformation on anorthosite can be locally observed where discrete zones of fine- to medium-grained gabbroic anorthosite grade from well-preserved igneous textures and minerals.

Return to Highway 101 and proceed east.

#### 75.2 STOP 3-5 - Thinly-layered tonalitic gneiss and diatreme breccia

Fine grained tonalitic gneiss at this exposure is strongly foliated and layered on a 1-5 mm scale with garnet, hornblende and biotite-rich layers. Extremely attenuated intrafolial folds are present locally. Units characterized by extremely planar foliation such as this are relatively rare in the Kapuskasing zone. Although the orientation of foliation in this exposure is typical for the Kapuskasing zone, most Kapuskasing gneisses are medium- to coarse-grained and layered with distinctive leucocratic portions. In addition, the layering in the typical gneisses is warped about gently northeast or southwest-plunging axes. The fine grain size and thin planar layering in this outcrop suggest a relatively late, high-strain flattening or shearing event.

A thin diatreme dyke occurs in this same exposure. It has not been dated but presumably is part of a set of lamprophyre dykes of ~1100 Ma age that occur in the Kapuskasing zone and are particularly common in the area between the Lackner and Nemegosenda Lake complexes. Both the matrix and fragments in the dyke are altered, but some fragments can be identified as massive pink granite. As massive granite does not occur in the Kapuskasing zone, the granite fragments are relatively exotic. Their source was probably below the Kapuskasing zone, possibly in granite of the Abitibi belt, which according to the gravity modelling, lies vertically below at a depth of ~15 km.



### 91.0 STOP 3-6 - Kapuskasing Gneiss

Layered mafic gneiss with tonalitic intrusions and sweats. Layering on the 5 to 10 cm scale is given by alternating hornblende-rich and pyroxene-rich layers. There are numerous minor folds with sheared limbs.

### 98.0 STOP 3-7 - Kapuskasing gneisses

There are several features of interest at this outcrop (Fig. 15):

A. Mafic gneiss is present on the northwest side of the road. It is a coarse grained rock consisting of three types of layers on the 5-100 mm scale: i) relatively anhydrous mafic rock made up of garnet, clinopyroxene, plagioclase and quartz, with some hornblende (analogous to analyses 1 & 3, Table 1); ii) more hydrous layers containing less garnet and clinopyroxene and more hornblende (analogous to analyses 2 & 4, Table 1); and iii) tonalitic leucosome layers, both concordant to layering and transverse in the amphibole-rich mafic rocks. Note that the tonalite has no retrogressive effect on adjacent anhydrous mafic gneiss. The tonalitic leucosome veinlets are considered to be in situ anatectic melt segregations developed during prograde metamorphic reactions (see reaction 2). In the western end of the outcrop, submicroscopic symplectites of orthopyroxene-plagioclase form barely-visible coronas around garnet, clinopyroxene and hornblende. Analyses of the symplectite minerals, at the lower size limit of microprobe resolution, are reported along with those from the other minerals in the rock, in Table 3. The rock contains three plagioclase compositions.  $An_{89}$  is present in coronas whereas worm-like intergrowths of  $An_{35}$  and  $An_{50}$  make up the matrix plagioclase.

The mineral compositions yield estimates of 735°C using the Ellis and Green (1979) garnet-clinopyroxene thermometer and 6.2 kbar using the garnet-clinopyroxene-plagioclase-quartz barometer (Newton and Perkins, 1982). At the same temperature the coronal minerals and matrix garnet yield 9.1 kbar with the garnet-orthopyroxene-plagioclase-quartz Newton and Perkins barometer.

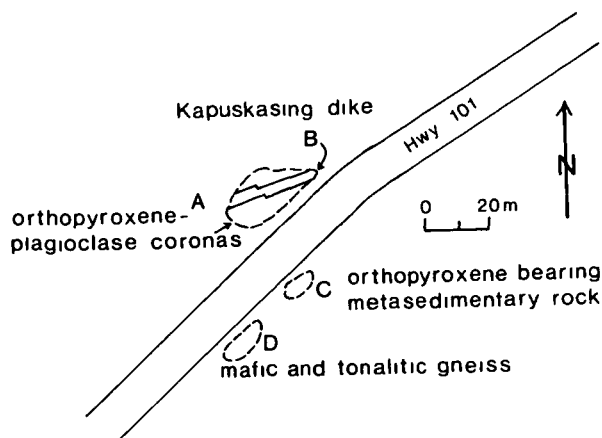


Fig. 15. Location of outcrops at Stop 3-7.

B. A Kapuskasing mafic dyke cuts the eastern end of the outcrop. The overall attitude of the dyke is 070/75 SE although the margin is offset by numerous small sinistral(?) faults. The outer 2 cm of the margin is chilled. Sparse plagioclase phenocrysts are present in the dominantly medium grained ophitic olivine-bearing gabbro. Several dykes of this swarm have been dated by the whole-rock K-Ar method and yield "ages" between 2367 and 3649 Ma, indicating the presence of excess argon (Stevens et al., 1981).

C. Homogeneous metasedimentary rock

South of the road is a flat outcrop of medium grained rock with the assemblage garnet-orthopyroxene-biotite-plagioclase-quartz. Plagioclase occurs as porphyroblasts to 2 cm and orthopyroxene is up to 5 mm. The rock has the same mineral assemblage as high-grade paragneiss in the Kapuskasing zone but lacks the migmatitic layering typical of paragneiss.

	1	2	3	4
148 SiO <sub>2</sub>	38.01	51.57	49.06	42.49
TiO <sub>2</sub>	0.00	0.34	0.03	2.03
Al <sub>2</sub> O <sub>3</sub>	20.99	2.92	1.75	12.98
Cr <sub>2</sub> O <sub>3</sub>	0.22	0.21	0.34	0.08
FeO*	28.06	11.81	31.20	18.43
MnO	0.70	0.00	0.81	0.17
MgO	4.11	11.34	13.35	9.28
CaO	8.32	22.65	1.39	11.41
Na <sub>2</sub> O	0.27	0.74	0.52	1.95
K <sub>2</sub> O	0.00	0.08	0.00	0.69
Total	100.67	101.79	101.45	99.31
Si	2.973	1.908	1.885	6.252
Al(iv)	0.00	0.092	0.115	1.748
Al(vi)	1.935	0.035	0.100	0.513
Ti	0.00	0.009	0.001	0.226
Cr	0.014	0.008	0.010	0.009
Fe <sup>3+</sup>	0.0	0.087	0.042	0.288
Fe <sup>2+</sup>	1.835	0.278	0.960	1.990
Mn	0.040	0.003	0.026	0.021
Mg	0.479	0.625	0.764	2.044
Ca	0.697	0.898	0.057	1.807
Na	0.041	0.053	0.039	0.559
K	0.00	0.004	0.00	0.130
(O)	(12)	(6)	(6)	(23)

1: garnet; 2: clinopyroxene, 3: orthopyroxene; 4: hornblende

\* Total iron as FeO; Fe<sup>3+</sup> by stoichiometry

Specimen also contains quartz and plagioclase (An<sub>33,57</sub>) in matrix, An<sub>88</sub> in symplectite)

Table 3. Microprobe analyses of minerals in coronitic mafic gneiss, stop 3-5.

#### D. Interlayered mafic and tonalitic gneiss

This outcrop demonstrates complex relations between mafic and tonalitic gneiss. Isoclinal folds of layering are truncated by tonalite pods and dykes, suggesting multiple generations of tonalite.

km

#### 107.2 STOP 3-8 - Ivanhoe Lake cataclastic zone

The outcrop south of the highway is on the western, high-grade side of the cataclastic zone and consists of migmatitic mafic gneiss with garnet-clinopyroxene-hornblende-plagioclase-quartz assemblages. It is transected by numerous small fault offsets and by one major cataclasite vein. In thin section, this black aphanitic material is seen to consist mainly of (recrystallized) fine actinolitic amphibole and of porphyroclasts of hornblende. A  $^{40}\text{Ar}/^{39}\text{Ar}$  whole-rock analysis of material from this vein yielded an age plateau at 1720 Ma (Fig. 13). On the west side of the outcrop are thin (3 cm) rusty-weathering lamprophyre dykes.

#### 108.7 STOP 3-9 - Mafic metavolcanics, Abitibi subprovince

This outcrop is on the eastern low-grade side of the ILCZ and is the westernmost exposure of metavolcanics of the Abitibi belt. It is a fine grained, layered hornblende-plagioclase  $\pm$  clinopyroxene rock with local rusty-weathering patches.

The structural characteristics of the outcrop, including east-west strike of layering, vertical dip and steeply-plunging isoclinal small folds, are typical of the Abitibi belt. Chlorite and epidote are common to the east along strike, where the belt is wider, suggesting an easterly decrease in metamorphic grade.

Analyses of hornblende and plagioclase from this outcrop by the  $^{40}\text{Ar}/^{39}\text{Ar}$  method (Fig. 13) show a plateau for hornblende at 2567 Ma and a saddle-shaped spectrum for plagioclase with a plateau at 1107 Ma. The plagioclase plateau may be due to argon loss resulting from a mild thermal event, possibly related to faulting.

Proceed east to Timmins.

#### DAY 4 - GEOLOGY OF THE ABITIBI GREENSTONE BELT, TIMMINS AREA

The main emphasis will be on the rock types and stratigraphy of the volcanic and sedimentary rocks of the Abitibi Supergroup. Stops will be confined to exposures of volcanic rocks of the Tisdale Group and sedimentary rocks of the Porcupine Group.

##### General Geology

With the exception of a few diabase dykes and Huronian sedimentary rocks of Proterozoic age, the bedrock of the area is of Archean age ( 2650 to 2750 Ma). The supracrustal rocks have been divided by Pyke (1982) into three groups, an older(?) metavolcanic Deloro Group, a younger metavolcanic Tisdale Group, and a metasedimentary Porcupine Group, considered to be stratigraphically equivalent to the upper part of the Deloro Group and the entire Tisdale Group (Fig. 16).

The Deloro Group is largely a calc-alkaline sequence, approximately 4500-5000 metres thick, and is composed mainly of andesite and basalt flows in the lower part and dacitic and rhyolitic flows and pyroclastics in the upper part. Iron formation is common at or near the top of the group. A major change in volcanic rock composition marks the lower contact of the Tisdale Group. The basal formation consists mainly of ultramafic volcanic rocks and basaltic komatiites. This is overlain by a thick sequence of tholeiitic basalts. The uppermost formation is largely volcaniclastic rocks of calc-alkaline dacitic composition. The total thickness of the Tisdale Group is almost 7000 metres.

The Porcupine Group consists of a lower turbiditic sequence of greywacke, siltstone and conglomerate, and an upper sequence of crossbedded fluvial sandstone and conglomerate. The total exposed thickness of the group is approximately 3000 metres.

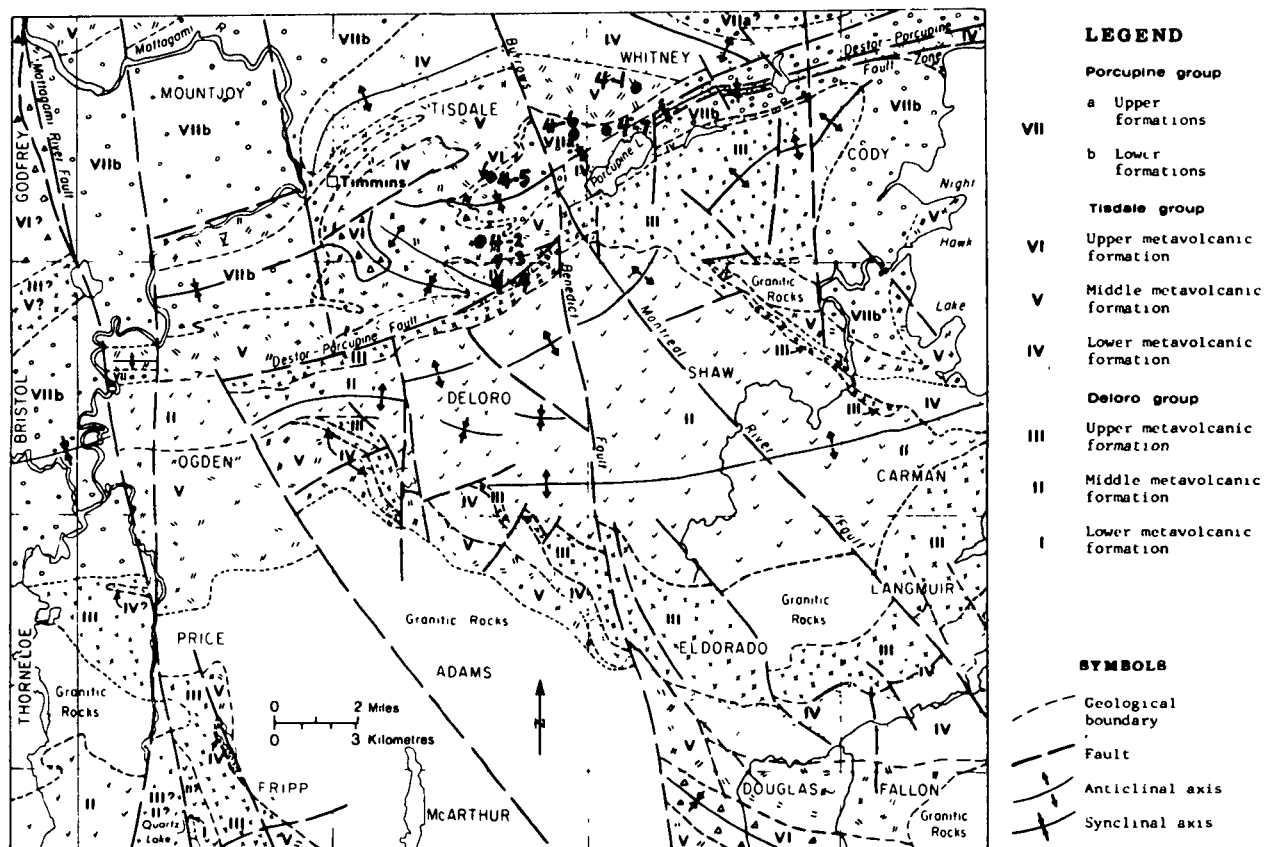


Fig. 16. Geology of the Timmins area (after Pyke et al., 1978).

Large sill-like bodies of dunite and lherzolite occur within the Deloro Group. Some of these bodies may be magma chambers for the overlying ultramafic volcanics of the Tisdale Group.

Small quartz-feldspar porphyry bodies are probably subvolcanic intrusions and may be part of intrusive-extrusive rhyolitic domes. Plutons and batholiths of trondhjemite, granodiorite, granite and syenite intrude the supracrustal rocks. Several of the massive, unmetamorphosed granodioritic intrusions yield U-Pb zircon ages close to 2680 Ma, thus approximately dating the termination of major orogenesis in this part of the Superior Province.

#### **STOP 4-1 - Mental Hospital Stop**

This stop is located just north of the Resource Centre near the northeast end of Porcupine Lake. Travelling eastward from Timmins, turn north off Highway 101 on the road just before the Resource Centre. Then turn west (0.9 km) on the first road and proceed 100 m to the start of the trail.

This stop illustrates the upper part of formation IV of the Tisdale Group and the overlying basal portion of formation V. The contact between formations IV and V is placed at the first readily recognizable iron-rich tholeiitic basalt. This Fe-tholeiite is possibly correlative with the 99 flow which underlies the V8 flow in Tisdale Township.

At this stop the upper part of formation IV consists of light grey Mg-tholeiitic basalt. That part of the lowermost flow included in the stop consists of pillow basalt and a thick sequence of pillow breccia. The overlying flow, also an Mg-tholeiite is massive at the base and pillowed in the upper part. Overlying this is a massive, medium grained, medium to dark green, Fe-tholeiitic basalt, which marks the base of formation V. This is overlain by variolitic pillow basalt, herein correlated with the V8 basalt in the Timmins gold camp. The overlying basalt is pillowed, locally contains very fine varioles and displays concentric cooling fractures. The overlying and uppermost flow on this stop is a

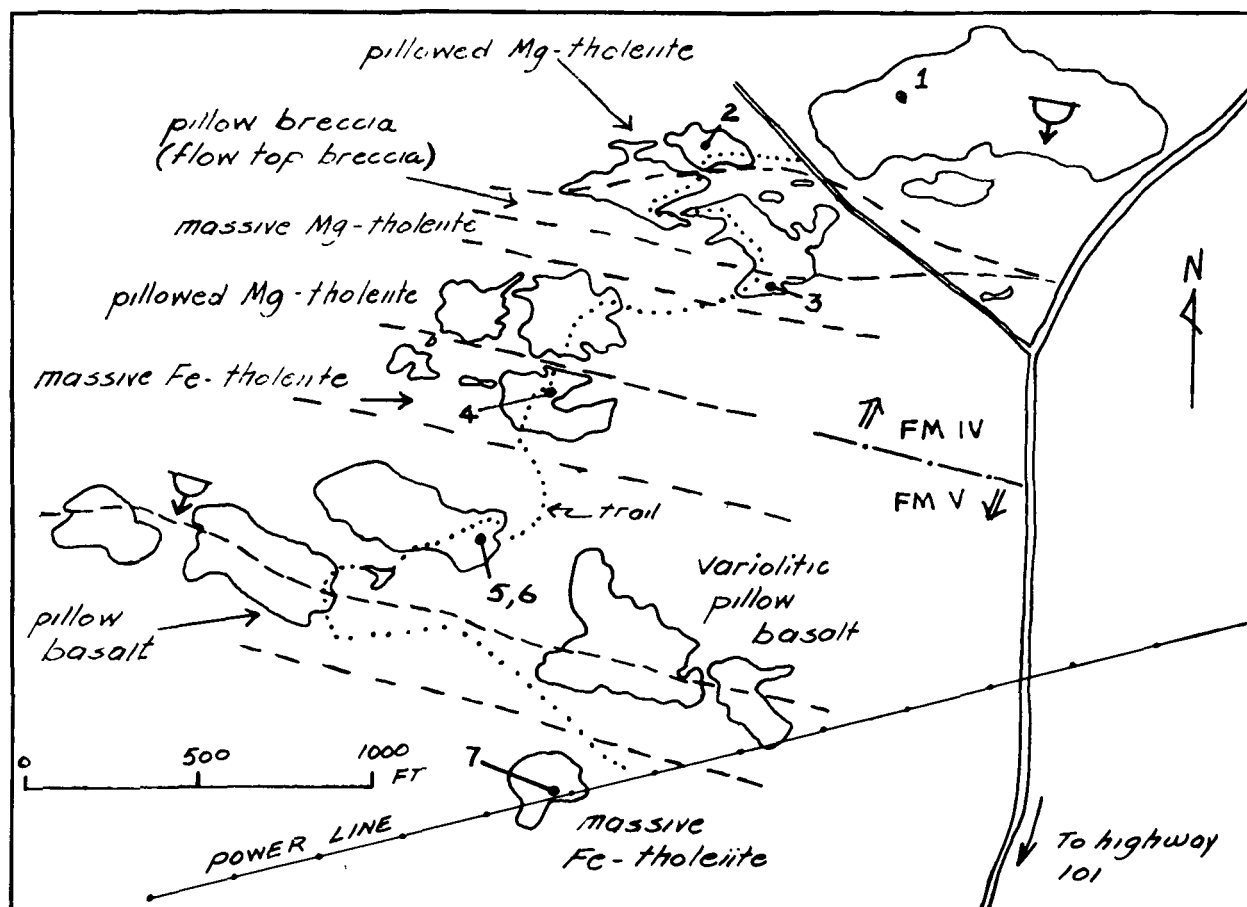


Fig. 17. Geology of Stop 4-1 (from Pyke et. al., 1978).

fine grained dark green Fe-tholeiitic basalt. Although Mg-tholeiites are interlayered with Fe-tholeiites in formation V in Whitney Township, they are absent in most of Tisdale Township.

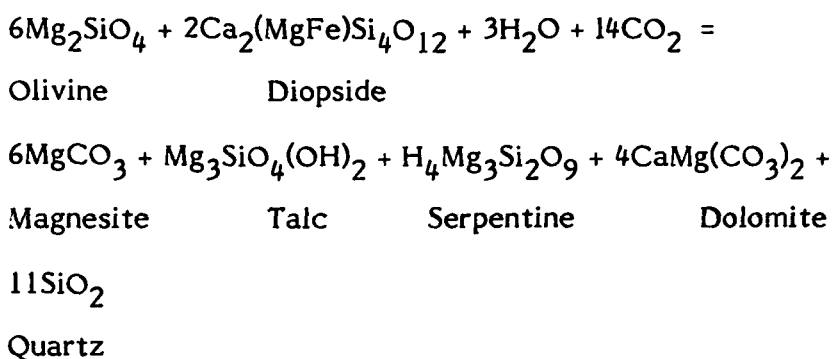
#### STOP 4-2 - Carbonated Ultramafic

This stop is on the back road between Timmins and South Porcupine just south of the Dome No. 3 head frame.



The outcrop is an altered peridotitic komatiite completely altered to carbonate minerals. Ultramafic flows such as this are almost wholly confined to the base of the Tisdale Group. The carbonatization has not destroyed the polysuturing which serves as an aid in the field to readily identify the rocks as being an ultramafic flow. This structure refers to the fracture pattern which is exhibited by the semi-equant polygons resembling mud cracks. Polysuturing is pervasive throughout ultramafic flows and probably represents some type of cooling phenomena, perhaps related to incipient pillow development.

A simplified equation for the alteration reaction of a komatiite might be:



The irregular quartz stringers throughout the rock may be due to the released free silica being deposited in fractures.

#### **STOP 4-3 - Paymaster Porphyry**

This stop is a few hundred metres east of the last stop.

The Paymaster porphyry is typical of quartz-feldspar porphyries of the area. It exhibits a light grey to buff weathering and is light grey to light green-grey on fresh surfaces. Quartz and albite crystals (phenocrysts or metacrysts?) are within a strongly foliated groundmass of fine albite, quartz and sericite.

The porphyries all occur within the lower formation of the Tisdale Group. Restriction to this stratigraphic interval suggest that they could represent rhyolitic

domes formed on the surface. This extrusive origin would explain the lack of offsets across contacts and the absence of evidence for forceful intrusion. If the porphyries within the Tisdale Group are not in part extrusive, there can be little doubt that they are high level intrusions of subvolcanic derivation.

#### **STOP 4-4 - Paymaster Mine**

This stop is at the Porcupine Paymaster Mine, opposite Simpson Lake on the back road connecting Timmins and South Porcupine.

The stop illustrates some of the iron-rich tholeiitic basalts typical of formation V of the Tisdale Group. The most diagnostic feature of the Fe-tholeiites is the dark green colour as compared to the Mg-tholeiites previously visited at Stop 1. The variolitic basalts exposed here are interpreted to correlate with the variolitic basalts of Stop 1, Whitney Township. They form an important stratigraphic marker throughout the gold camp. At the Paymaster, a relatively dark green, vesicular, pillowed Fe-tholeiitic basalt (1) is exposed on the south side of the highway. Minor varioles are present and are a common feature to many of the flows in this part of the section. This pillowed flow also outcrops immediately north of the highway, as does a poorly exposed variolitic basalt (2). This in turn is overlain by an Fe-tholeiitic basalt (3) which is dominantly massive and contains minor blue quartz. This is in sharp contact with a variolitic pillow basalt. The pillows are commonly large (2-3.6 m in length) and the varioles are up to 4-5 cm in diameter, and may coalesce to form patches up to 45 cm in maximum dimension. Minor massive sections can be traced out within the dominantly pillowed variolitic basalt, and may represent intercalated flows. The coarse variolitic flow is overlain by a massive basalt (5) and this in turn by a variolitic pillow basalt (6). These flows are succeeded to the north by at least two and possibly four massive Fe-tholeiitic basalts. The northernmost basalt is highly carbonatized.

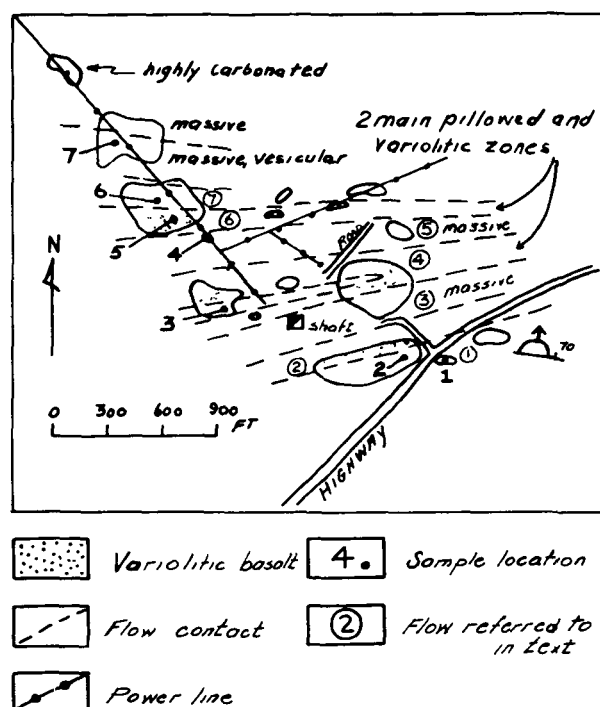


Fig. 18. Geology of Stop 4-4 (from Pyke et al., 1978).

#### STOP 4-5 - Krist Fragmental

This stop is approximately midway between Schumacher and South Porcupine, and is located at the large outcrop area adjacent to the south side of Highway 101 where the highway intersects a north-trending power line.

The Krist fragmental is a pyroclastic tuff breccia which forms the upper part (formation VI) of the Tisdale Group. The breccia is white-weathering, massive, and contains fragments of crystal tuff in a matrix of similar composition. The crystal fragments are dominantly white-weathering, subhedral, albitic feldspar averaging 1.5-2.5 mm in size; minor (5-10 per cent) quartz of comparable size is also present.

Although not readily obvious, the fragments are lineated and plunge eastward at approximately 45 degrees. Crystal tuff outcrops immediately adjacent to the highway, and rapidly passes into tuff-breccia farther down the power line. Most breccia fragments range from 2.5-15 cm in maximum dimension; the largest are about 45 cm. The Krist fragmental perhaps represents a glowing avalanche type of volcanic deposit.

#### **STOP 4-6 - Unconformity Outcrop**

This stop is located 1.0 km north from Highway 101 along Crawford Street. Turn north at the Esso Service Station at the northeast end of South Porcupine. A trail west is just past the line between Concessions III and IV, Tisdale Township.

This stop illustrates the unconformity between what has classically been termed the "Keewatin" and "Timiskaming" sedimentary rocks in the Timmins area (Ferguson et al., 1968). The older "Keewatin" sediments, turbiditic wacke and siltstone, strike about 105°, dip steeply north, and face south conformably with the underlying volcanic rocks. Good examples of scouring, rip-up clasts, load casts and a Bouma sequence (A to D) may be seen in the northernmost exposures. A poorly sorted polymictic conglomerate approximately 60 m thick, trends 065 degrees and marks the base of the unconformity. Boulders are predominantly of basalt, siltstone, greywacke, and lesser gabbro and felsic volcanics. There does not appear to be any tectonic unconformity between the "Keewatin" and "Timiskaming sediments" as the structures within the two sedimentary sequences are the same.

#### **STOP 4-7 - Shovel Outcrop**

This stop is at the first outcrop on Government Road, east of the turnoff to Stop 6.

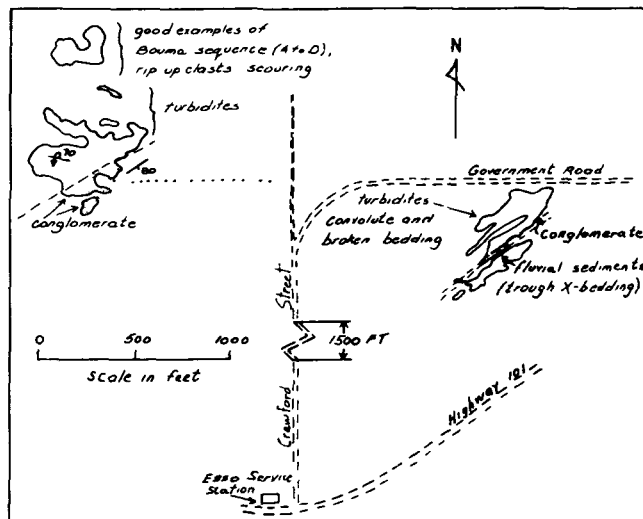


Fig. 19. Geology of Stops 4-6 and 4-7 (after Pyke et al., 1978).

The sediments at this stop perhaps contain the most significant sedimentary contact in the Timmins area. Here, there is a major change from turbidite to fluvial type of sedimentation. The contact between the two types of sedimentary rocks is not exposed, but the bedding is conformable. At the top of the turbidite sequence (base of the fluvial sediments?) is a narrow (1.8-2.4 m) polymictic conglomerate, conformable with the turbidites. Convolute bedding and disrupted and broken beds are common in the turbidites. The fluvial sediments are quartz-rich sandstones (lithic arenites) and display abundant trough crossbedding. This contact may warrant close scrutiny for gold mineralization.

## Analysis (Wt%)

159

	7	8	9	10
SiO <sub>2</sub>	54.7	43.9	72.7	55.6
Al <sub>2</sub> O <sub>3</sub>	13.2	7.18	13.9	12.5
Fe <sub>2</sub> O <sub>3</sub>	10.3	9.28	1.20	8.87
FeO	0.0	0.0	0.0	0.0
MgO	9.56	13.7	0.40	10.7
CaO	4.35	14.7	1.55	3.21
Na <sub>2</sub> O	3.02	0.61	6.50	1.76
K <sub>2</sub> O	0.62	1.77	1.25	0.50
TiO <sub>2</sub>	0.80	1.00	0.04	0.65
P <sub>2</sub> O <sub>5</sub>	0.11	1.80	0.02	0.07
S	0.0	0.0	0.0	0.0
MnO	0.18	0.20	0.23	0.12
CO <sub>2</sub>	0.0	0.0	0.0	0.0
H <sub>2</sub> O <sup>+</sup>	3.50	6.60	1.50	6.60
H <sub>2</sub> O <sup>-</sup>	0.0	0.0	0.0	0.0

Total	100.3	100.7	99.3	100.6
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Cu	74	192	20	66
Zn	90	81	64	118
Ni	340	73	24	620
Cr	840	236	32	840

- 7. Conglomerate
- 8. Basaltic komatiite
- 9. Calc-alkalic rhyolite
- 10. Argillite

## NORMS MOLECULAR

AP	0.238	4.013	0.042	0.157
PO	0.0	0.0	0.0	0.0
IL	1.152	1.483	0.056	0.967
OR	3.792	11.146	7.482	3.158
AB	28.037	5.831	59.060	16.876
AN	21.339	12.376	5.122	16.522
C	0.0	0.0	0.0	3.959
AC	0.0	0.0	0.0	0.0
MT	2.487	2.783	0.463	2.401
HM	0.0	0.0	0.538	0.0
WO	0.0	0.0	0.0	0.0
EN	27.220	1.705	0.106	31.548
FS	8.990	0.299	0.0	7.638
Q	6.556	0.0	25.106	16.776
DI	0.142	35.900	2.023	0.0
FO	0.0	15.461	0.0	0.0
FA	0.0	2.710	0.0	0.0
NE	0.0	0.0	0.0	0.0
LC	0.0	0.0	0.0	0.0
KP	0.0	0.0	0.0	0.0
HE	0.047	6.293	0.0	0.0
CC	0.0	0.0	0.0	0.0
RU	0.0	0.0	0.0	0.0
NS	0.0	0.0	0.0	0.0
KS	0.0	0.0	0.0	0.0
CR	0.0	0.0	0.0	0.0
LN	0.0	0.0	0.0	0.0

## DAY 5 - KOMATIITES AND SEDIMENTS OF THE ABITIBI BELT IN THE KIRKLAND LAKE AREA

On the return trip to Ottawa we will visit an outcrop area in the Abitibi belt southeast of Kirkland Lake. The aim of the stop will be to show the relationship of komatiites in Archean volcanic sequences and their relationships to other rock types.

In the Abitibi belt, komatiites form the initial phases of several major volcanic cycles. Each cycle consists of a lower komatiitic sequence followed in turn by tholeiitic, calc-alkalic, and, locally, alkalic phases. Two such cycles are present in the Kirkland Lake area. At this stop komatiites of the second cycle overlie rhyolites of the first cycle, indicating that these komatiites are not primitive crust nor were they deposited on primitive crust. It is possible that in the central parts of the basin komatiitic lavas were deposited directly on primitive Archean crust, but at this locality they overlap a pre-existing volcanic pile. Erosion of this volcanic pile occurred, and hence, turbiditic sediments with detritus from both the komatiites and rhyolites are interbedded with the volcanic rocks.

### STOP 5-1

Stop 1 shows pyroxene spinifex toward the top of a basaltic komatiite flow (analysis 8). Overlying the flow is massive calc-alkaline rhyolite tuff breccia(?). An amygdaloidal dyke cuts both the basaltic komatiite and the calc-alkalic rhyolite.

### STOP 5-2

Stop 2 shows the rhyolite tuff (analysis 9) interlayered with conglomerate composed mainly of ultramafic clasts. The conglomerates contain clasts of ultramafic and basaltic komatiite and magnesium-rich tholeiite, i.e. all the mafic volcanic rock-types found in the area. Clasts with olivine and pyroxene spinifex textures are present in the conglomerate. A large rhyolite-ultramafic komatiite boulder 3 x 2 m in size, occurs in the upper conglomerate. It may be noted mafic volcanic clasts are absent in the rhyolite tuff.

Many of the ultramafic conglomerates are difficult to distinguish from polysutured ultramafic flows and where slightly sheared and altered, are indistinguishable unless isolated spinifex clasts or clasts of rhyolite or trachyte can be observed.

#### STOP 5-3

Stop 3 shows a small peridotite intrusion cutting the sedimentary rocks. It is one of many found in the basinal area. Its texture and appearance are quite distinctive from the ultramafic komatiites.

#### STOP 5-4

Stop 4 shows a finely bedded turbiditic sequence. Flame structures and many other features associated with such rocks can be observed at this stop.

Some beds of the outcrop were disturbed during their deposition.

The chemical composition of the argillite (analysis 10) has the composition of basaltic komatiite. Analysis 7, is of a similar sedimentary rock found lower in the sequence.

#### STOPS 5-5 and 6

Stops 5 and 6 are coarser grained turbidites which have graded bedding, channel scouring; laminations can be seen at the tops of many beds. At stop 6, the sedimentary rocks range from boulder conglomerate to fine grained argillite.



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# List of Invited Participants

Lewis D. Ashwal  
*Lunar and Planetary Institute and Yale University*

Gary P. Beakhouse  
*Ontario Geological Survey*

Kenneth Card  
*Geological Survey of Canada*

Richard Carlson  
*Carnegie Institution of Washington*

Ken Collerson  
*Australian National University*

Kent C. Condie  
*New Mexico Institute of Mining and Technology*

R. F. Dymek  
*Harvard University*

Richard Ernst  
*University of Toronto*

Norm Evensen  
*University of Toronto*

Tomas Feininger  
*Department of Energy, Mines and Resources, Canada*

Robert F. Folinsbee  
*University of Alberta*

David M. Fountain  
*University of Wyoming*

R. A. Frith  
*Geological Survey of Canada*

Gilbert N. Hanson  
*State University of New York, Stony Brook*

Larry A. Haskin  
*Washington University*

Darrell Henry  
*Lunar and Planetary Institute*

Lincoln S. Hollister  
*Princeton University*

Larry Jensen  
*Ontario Geological Survey*

Cornelis Klein  
*Indiana University*

L. E. Luhta  
*Ontario Ministry of Natural Resources*

Dr. Frank R. Luther  
*University of Wisconsin, Whitewater*

J. D. Macdougall  
*Scripps Institute of Oceanography*

George E. McGill  
*University of Massachusetts*

Paul A. Mueller  
*University of Florida*

Elaine R. Padovani  
*National Science Foundation*

James J. Papike  
*South Dakota School of Mines and Technology*

John Percival  
*Geological Survey of Canada*

William Phinney  
*NASA Johnson Space Center*

Ullrich Rast  
*Max-Planck-Institut für Chemie*

Roberta Rudnick  
*Australian National University*

R. P. Sage  
*Geological Survey of Canada*

Peter Salpas  
*Washington University*

Craig Schiffries  
*Harvard University*

Klaus J. Schulz  
*U.S. Geological Survey*

Denis M. Shaw  
*McMaster University*

Steven B. Shirey  
*State University of New York, Stony Brook*

Richard Stern  
*Geological Survey of Canada*

Greg M. Stott  
*Ontario Geological Survey*

Paul Sylvester  
*Washington University*

S. Ross Taylor  
*Australian National University*

George R. Tilton  
*University of California, Santa Barbara*

Marilyn Truscott  
*McMaster University*

John W. Valley  
*University of Wisconsin, Madison*

Werner Weber  
*Mineral Resources Division, Manitoba*

J. F. Wilson  
*University of Zimbabwe*

Xianglin Qian  
*Peking University and University of Toronto*