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# PRECAMBRIAN GEOLOGY OF THE SOUTHERN CANADIAN SHIELD AND THE EASTERN BALTIC SHIELD

U.S.A.–U.S.S.R.–Canada Joint Seminar,
August 21–23, 1990, Duluth, Minnesota

UNIVERSITY OF MINNESOTA

Minnesota Geological Survey
*Priscilla C. Grew, Director*

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Edited by

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University of Minnesota, Duluth

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# EDITOR'S FOREWORD

The geologic histories of the Canadian and Baltic Shields in North America and Europe, respectively, are broadly similar, and the topic was discussed during a conference and field trip involving North American and Russian participants in the late summer of 1990. During a two-day meeting prior to the field trip, twelve North American and eleven Soviet geologists presented papers, and participants discussed a variety of problems and ideas in Precambrian stratigraphy, sedimentology, tectonics, magmatism, industrial minerals, and metallogeny. Special emphasis was placed on problems of correlation. All papers were simultaneously interpreted by Senior Translator and Interpreter Grigori Sokolov of the Institute of Geology, Karelian Branch, U.S.S.R. Academy of Sciences, who accompanied the Russian delegation. His ability contributed greatly to the meeting's success. In addition to the speakers, thirty-eight geologists attended the conference: four Canadians, two Finns, and thirty-two Americans, including eight graduate students.

As a result of the seminar and field trip, exciting and promising opportunities for continued cooperation were identified. Specific proposed activities include meetings, field excursions, short courses, joint publications, individual research-oriented exchanges, and joint projects. Involvement of young geologists was especially encouraged to promote long-term cooperative relationships. Opportunities also were identified for cooperation with other international projects, such as existing bilateral programs and the International Geological Correlation Program.

It was mutually agreed that in 1991-1992, the Institute of Geology, Karelian Research Center, and the Kola Research Center of the USSR Academy of Sciences will host conferences and field trips on Proterozoic and Archean geology and metallogeny in the eastern Baltic shield. In 1991, the field program will emphasize Proterozoic geology, and in 1992, Archean geology. Other joint activities in the future will depend on the outcome of the 1991 and 1992 meetings.

It was the intent of the organizers to bring this joint activity to the attention of officials involved in relevant international programs. Toward that end this proceedings volume has been published by the Minnesota Geological Survey.

The body of this report consists of two parts; the first is a series of short papers that provide an overview of the Precambrian geology in the Great Lakes Region; the second part consists of a similar overview of the eastern part of the Baltic Shield.

North American participants were asked to submit extended abstracts that could be collated and distributed at the time of the meeting. Those contributions appear in this volume for the most part as they were received from the authors. It also was planned to distribute the contributions of the Russian participants at the time of the meeting, but several subsequent events delayed distribution until after the meeting. These papers were somewhat revised and edited after the meeting, and the revised versions are included herein. Avis Hedin and Joan Hendershot assisted in the laborious process of retyping successive revisions. Mary Nash, Executive Secretary of the Department of Geology at Duluth, provided outstanding administrative support.

All of the conveners, Priscilla Grew and G.B. Morey of the Minnesota Geological Survey, K.D. Card of the Geological Survey of Canada, and R.W. Ojakangas of the University of Minnesota-Duluth, wish to thank the following organizations for their support of the 1990 meeting: U.S. National Science Foundation, U.S.S.R. Academy of Sciences, Geological Survey of Canada, University of Minnesota-Duluth Department of Geology, Minnesota Geological Survey, Ontario Geological Survey, and the Soros Foundation - Soviet Union.

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# GENERAL COMPARISON OF THE CANADIAN AND BALTIC SHIELDS

K.D. Card, Geological Survey of Canada, Ottawa

| | Superior | Kola |
|---|---|---|
| *Age span* | 3.1-2.6 Ga | 3.1-2.6 Ga |
| *Major orogenies* | ca. 2.9 Ga - "Wanipigowan" | 3.1-2.9 Ga - Saamian |
| | ca. 2.7 Ga - Kenoran | 2.9-2.7 Ga - Lopian (Rebolian) |
| *Major lithotectonic units* | 1) 3.1-2.8 Ga tonalitic-granodioritic plutonic rocks with greenstone remnants forming early sialic crustal elements | 1) 3.1-2.9 Ga tonalitic-granodioritic plutonic rocks with greenstone remnants forming early sialic crustal elements |
| | 2) 3.0-2.8 Ga platformal sequences (quartz arenite, stromatolitic marble, mafic ultramafic vols) unconformable on older plutonics | 2) 2.9-2.7 Ga Kola Peninsula gneisses? |
| | 3) 2.8-2.7 Ga greenstone belt sequences; submarine plain mafic/ultramafic tholeiitic-komatiitic sequences; central volcanic complex, tholeiitic-calcalkalic sequences and synvolcanic plutons, Timiskaming-type alluvial sediment-shoshonitic/alkalic volcanic sequences | 3) 2.9-2.6 greenstone belt sequences; komatiitic, tholeiitic, and calcalkalic volcanics and syn-volcanic plutons |
| | 4) ca 2.7 Ga low- to high-grade metasedimentary sequences | 4) 2.9-2.6 Ga high-grade metasedimentary sequences |
| | 5) 2.7-2.6 Ga plutonic suites; mafic-felsic (gabbro to granite), syn- to post-tectonic re Kenoran orogeny | 5) 2.7-2.6 Ga plutonic suites; syn- to post-tectonic re Lopian orogeny (Chupa cycle) |
| | 6) Granulite gneisses uplifted in the late Archean or Early Proterozoic | 6) Granulite gneisses uplifted in the Early Proterozoic |
| *Major terranes and their characteristics* | 1) Older (3.1-2.8 Ga) crustal elements of Sachigo, Uchi, and Wabigoon subprovinces; tonalitic granodiorite gneiss with greenstone remnants and local platformal cover sequences | 1) Older Saamian (3.1-2.9 Ga) crustal elements of Karelian province, tonalite-granodiorite gneiss with greenstone remnants |
| | 2) Younger (2.8-2.7 Ga) granite-greenstone subprovinces, e.g. Abitibi, low-to medium-grade, low P metamorphism, polyphase deformation | 2) Younger (2.9-2.7 Ga) granite-greenstone terranes, i.e. Karelian belts, low- to medium-grade, low P metamorphism, polyphase deformation |

| | | |
|---|---|---|
| | 3) Metasedimentary gneiss belts, e.g. Quetico, low- to high-grade metamorphism, abundant granitic intrusions, polyphase deformation including early thrusts, recumbent folds | 3) Linear belts of amphibolite-granulite facies metasedimentary gneiss, e.g. Belomorian, early recumbent folding; multiple medium to high pressure meta accompanying plutonism |
| | 4) Granulite terranes uplifted during Early Proterozoic collisional events, e.g. Kapuskasing structural zone | 4) Granulite terrane uplifted during Early Proterozoic collisional event; i.e. Lapland granulite belt |
| *Late Archean-Early Proterozoic history* | Craton formation and stabilization during 2.7-Ga Kenoran orogeny followed by rifting and deposition of Huronian volcanic and sedimentary sequences ca. 2.5-2.4 Ga ago | Craton formation and stabilization during 2.7-Ga Lopian orogeny followed by rifting and deposition of Lapponian volcanic and sedimentary sequences ca 2.6-2.3 Ga ago |
| *Tectonic interpretation* | Subduction-driven, southward-younging accretion of volcanic arc, back-arc, and sedimentary prisms accompanied and followed by plutonism and transpression (Card, 1990) | Subduction-driven, westward-younging accretion of volcanic arc, back-arc, and sedimentary prisms accompanied and followed by plutonism (Gaal and Gorbatschev, 1987) |

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# Precambrian geology of the Lake Superior region-- An overview

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## Geologic framework

The Lake Superior region lies along the southern exposed part of the Canadian Shield, and because of the long duration and intensity of study by several generations of geologists is considered the classic area of Precambrian rocks in the United States. The Precambrian rocks record an interval of crustal development that spans nearly 3 b.y. of earth history. This interval of geologic time is not continuously recorded in layered and intrusive units, but instead is punctuated by specific rock-forming and tectonic events that can be deduced from geologic relations and placed in a chronometric framework by isotopic dating (Morey and Van Schmus, 1986).

The Precambrian rocks in the Lake Superior region constitute the southern part of the Archean Superior craton, an Early Proterozoic orogenic belt (Penokean orogen; 2.0-1.83 Ga) that marginally affects the Archean craton, and a middle Proterozoic intracratonic rift assemblage (~1.1 Ga), as well as scattered Middle Proterozoic intracratonic igneous and sedimentary rocks (fig. 1).

## Archean terranes

The Archean is a composite crustal segment consisting of a Late Archean greenstone-granite terrane and an Early to Late Archean gneiss terrane (Morey and Sims, 1976), which are juxtaposed along a major tectonic boundary named the Great Lakes tectonic zone (Sims and others, 1980). The greenstone-granite terrane, the southernmost part (Wawa subprovince) of the Superior province of Canada, occupies northern Minnesota and adjacent Canada and parts of northern Michigan and Wisconsin (fig. 1). The Wawa subprovince, in Minnesota and Michigan-Wisconsin, is juxtaposed against the Archean gneiss terrane; it is separated from the Quetico subprovince on the north by the Vermilion fault. The Quetico subprovince in turn is separated from the Wabigoon subprovince by the Rainy Lake-Seine River and Quetico faults.

The Wawa terrane consists of volcanic and sedimentary rocks and subvolcanic plutons, deformed by late upright, plunging folds imposed on early tight recumbent folds (Hudleston and others, 1988; Bauer, 1985), metamorphosed to greenschist and, locally, amphibolite facies, and cut by post-tectonic, calc-alkaline to alkaline 2.68 Ga granite bodies. The late stage of deformation involves rocks as young as 2.69 Ga and resulted from dextral transpression (Poulsen, 1986; Hudleston and others, 1988). The volcanic rocks are bimodal, consisting of tholeiitic and less commonly komatiitic basalt intercalated with iron-formation and calc-alkaline dacite-rhyodacite (Schulz, 1980), 2.75 to 2.71 Ga, intruded by coeval tonalitic plutons. Locally, conglomerates and other terrestrial and marine epiclastic rocks (Temiskaming-type) unconformably overlie the older rocks (Thurston and Chivers, in press); they were deposited about 2.69 Ga (Corfu and Stott, 1986).

The Quetico subprovince, in Minnesota, consists of the Vermilion Granitic Complex (Southwick and Sims, 1980)--mainly metamorphosed and deformed turbidites of mixed felsic-mafic provenance--intruded variously by the Lac La Croix Granite (~2.65 Ga). Metamorphism is amphibolite grade. The granite grades outward from granite-rich migmatite through a narrow zone of schist-rich migmatite into biotite schist. Bauer (1985) has shown that an $F_1$ recumbent fold and an upright $F_2$ antiform can be correlated across the boundary between the Quetico migmatitic terrane and the adjacent Vermilion district (Wawa subprovince), indicating that the two disparate terranes were deformed together. Further, biotite schist of the Quetico subprovince is separated from greenschist-facies metagraywacke of the Vermilion district by a high-angle dip-slip fault (Haley fault; Sims and Southwick, 1985), suggesting that the biotite schist is a deeper seated equivalent of the metagraywacke. If this is correct, the Haley fault represents a metamorphic boundary rather than an accretionary boundary. In contrast, in adjacent Canada, the Quetico prism is interpreted as having been constructed of trench turbidites accreted to the Wabigoon (fore arc) subprovince during dextral-oblique northward subduction terminated by collision with the back of the Wawa arc (Percival and Williams, 1989).

Gneisses of the Archean gneiss terrane (Morey and Sims, 1976) are exposed in the Minnesota River Valley, where erosion has cut below a thin cover of Cretaceous sedimentary rocks, and in east-central Minnesota and northern Michigan, where they mainly compose the cores of mantled gneiss domes or uplifted fault blocks (Morey and others, 1982). These rocks probably formed a continuous crust (protocontinent) of wide areal extent prior to Middle Proterozoic (Keweenawan) rifting. Archean gneisses exposed in central Wisconsin (Marshfield terrane, fig. 1) probably are not contiguous with the larger Archean crustal segment, and accordingly they are not considered as part of the Archean gneiss terrane. In the Minnesota River Valley, the gneiss terrane is characterized (see Sims and Peterman, 1981, and references therein) by migmatitic gneisses and amphibolite 3,000 Ma or more old, amphibolite- and granulite-facies metamorphism, and generally moderately open folding. At three localities, 3,500 Ma ages have been determined on the gneisses. The gneisses were modified by addition of a neosome about 3,000 Ma, which was followed by folding and high-grade metamorphism, and by intrusion of post-tectonic granite at ≈2,600 Ma, after a second episode of tectonism.

The two major Archean terranes largely evolved separately. They were juxtaposed in the Late Archean (≈2,690 Ma) along the Great Lakes tectonic zone (Sims and others, 1980) by collision of the gneiss terrane with the southern margin of the Superior province (greenstone-granite terrane). The collision was oblique and resulted in dextral-thrust shear along the boundary, northwestward vergence, and probable overriding of the greenstone-granite terrane by the gneiss terrane (Sims, 1990). Transmittal of the dextral shear stress across a large area of the Superior province crust to the north may have been responsible for the east-west foliation, upright folds, and northwest- to east-trending dextral faults and shear zones at least as far north as the Quetico fault (Qf, fig. 1). The shear zones host most of the Archean gold deposits in the southern part of the Superior province (Poulsen, 1983).

## Early Proterozoic Penokean orogen

The 2.0-1.83 Ga Early Proterozoic Penokean orogen is the zone of deformed and metamorphosed Early Proterozoic and Archean rocks along the southern margin of the Superior Archean craton. Rocks of the orogen are exposed in Minnesota, Wisconsin, and northern (Upper Peninsula) Michigan. The orogen consists of two distinct lithologic domains (fig. 1): (1) a northern deformed continental-margin prism overlying Archean basement, and (2) a southern assemblage of island arcs, the Wisconsin magmatic terranes (Sims and others, 1989). The northern rocks include the iron-bearing (Sims, 1987) Marquette Range Supergroup in Michigan and Wisconsin (Cannon and Gair, 1970) and the Animikie and Mille Lacs Groups in Minnesota (Morey, 1983). They consist of a lower rifted passive-margin sequence overstepped northward by a synorogenic foredeep sequence (Barovich and others, 1989; Southwick and others, 1988). Early Proterozoic deformation involved northward-directed thrusting and related folding of the supracrustal rocks and produced metamorphism and basement gneiss domes caused by crustal thickening (Klasner and others, 1988; Holm and others, 1988). The southern, magmatic terranes consist of Early Proterozoic calc-alkaline and tholeiitic volcanic and calc-alkaline plutonic rocks (Sims and others, 1989). The Niagara fault (suture) zone (≈1,850 Ma) separates the south-facing continental margin and the arc magmatic terranes in Wisconsin and Michigan. D.L. Southwick and G.B. Morey (written comm., 1990) propose that the Malmo discontinuity in east-central Minnesota could be the western extension of the Niagara fault zone (fig. 1).

Two distinctive terranes are distinguished within the Wisconsin magmatic terranes on the bases of lithology and structure. They are separated by the Eau Pleine shear zone, a presumed south-verging paleosuture (fig. 1). The northern (Pembine-Wausau) terrane apparently lacks Archean basement; the southern (Marshfield) terrane has an Archean basement. After amalgamation of the two terranes, at about 1,840 Ma, alkali-feldspar granite (≈1,835 Ma) was intruded as stitching plutons, and cogenetic silicic rhyolite was erupted from possible calderas in the vicinity of the Eau Pleine shear zone. The Pembine-Wausau terrane hosts important massive sulfide deposits (Mudrey, 1979; Sims, 1987).

## Middle Proterozoic Midcontinent rift system

The youngest major terrane in the Lake Superior region, the Middle Proterozoic (≈1.1 Ga) Midcontinent rift system, is an intracontinental assemblage of igneous and sedimentary rocks that formed in a rift that aborted before significant crustal separation was achieved (Wold and Hinze, 1982; Van Schmus and Hinze, 1985). The rocks are dominantly bimodal basalt and rhyolite, which occupy a central uplifted graben bounded by high-angle reverse faults; gabbro-anorthosite complexes, which were intruded along the unconformity along the margins of the rift between older Archean and Early Proterozoic rocks and the younger Keweenawan lava flows; and mainly red-bed, but locally carbon-rich, sedimentary rocks. Seismic reflection profiling has shown that the volcanic rocks and postvolcanic and interbedded sedimentary rocks extend to depths as great as 32 km; this may be the greatest thickness of intracratonic rift deposits on Earth (Behrendt and others, 1988). The seismic profiling is interpreted (Cannon and others, 1989) to indicate that the central graben is asymmetrical, and that, in addition to crustal sagging documented by previous investigations, normal faulting was important in subsidence of the axial region of the rift.

Near the axis, the prerift crust is thinned to about one fourth of its original thickness, apparently by a combination of low-angle extensional faulting and ductile stretching or distributed shear. In Michigan, the volcanic rocks and the sedimentary rocks host major copper ore bodies (White, 1968; Ensign and others, 1968).

## Tectonic evolution

In summary, suturing of the Archean gneiss terrane and the southern part of the Archean Superior province about 2.69 Ga produced dextral transpression (Hudleston and others, 1988) across the Archean greenstone-granite rocks in north-central United States. Following stabilization of the composite Archean crust, Early Proterozoic epicratonic rocks were deposited (≈2.0-1.9 Ga) on Archean basement along the rifted continental margin. The breakup of the continent led to ocean spreading, formation of volcanic arc systems from about 1.9 to 1.84 Ga (Wisconsin magmatic terranes), southward subduction along the southern rifted margin and, eventually, collision of the volcanic arcs with the passive margin (≈1.85 Ma; Penokean orogeny). The Penokean orogeny and its predominantly calc-alkalic magmatism was followed in the region by a 1.76 Ga episode of anorogenic activity (Sims and others, 1987). The youngest major tectonic event in the region was the development of the Midcontinent rift system and its coeval igneous and sedimentary rocks. One consequence of the rifting was development of the Goodman swell in northeastern Wisconsin, on the southern side of the rift; it is interpreted as a forebulge imposed on an elastic crust by loading of predominantly mafic igneous rocks along the axis of the Midcontinent rift system (Peterman and Sims, 1988).

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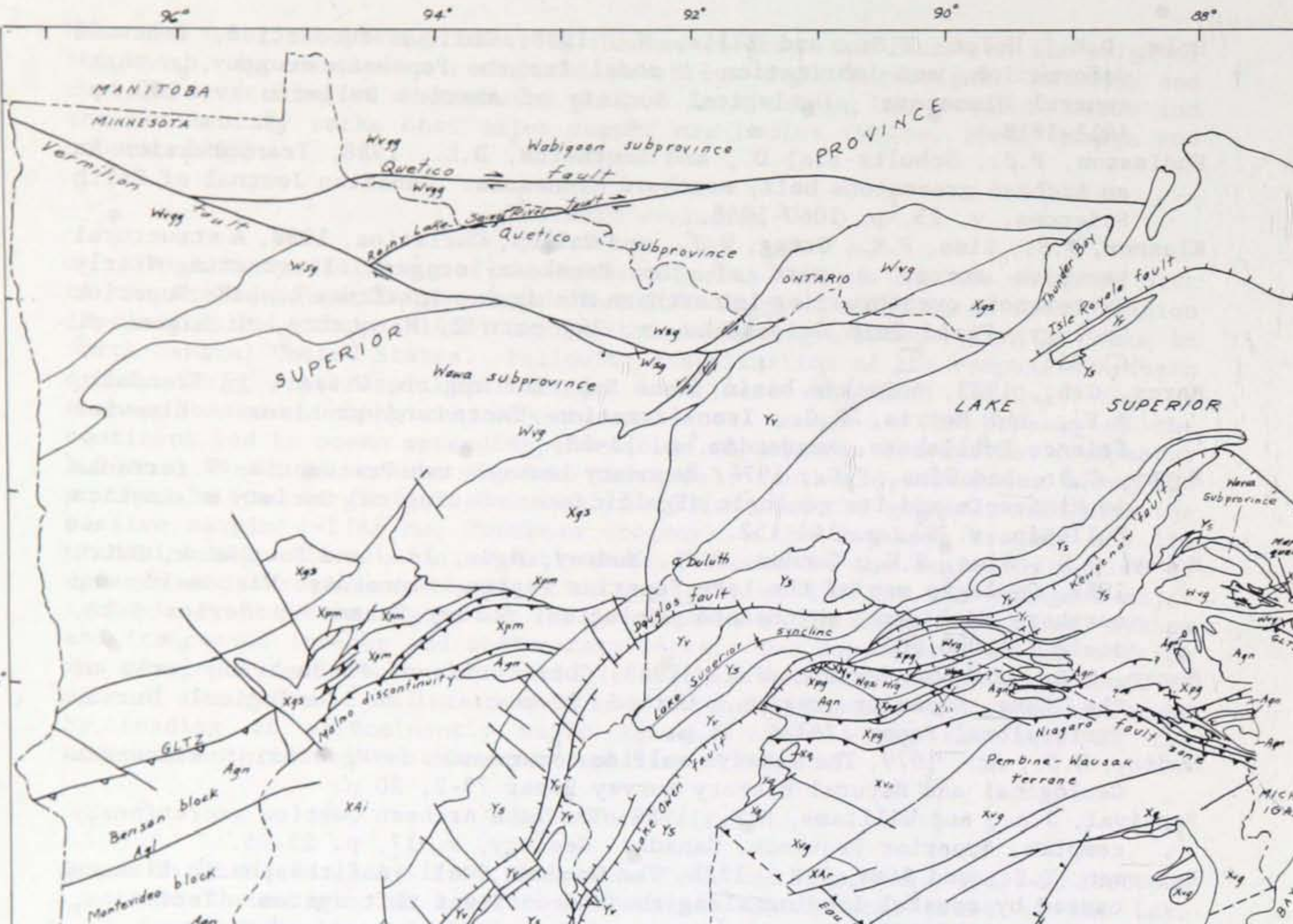
96°
94°
92°
90°
88°
48°
46°
44°
MANITOBA
MINNESOTA
Vermilion fault
Quetico
fault
Wabigoon subprovince
PROVINCE
Rainy Lake
Seine River
Quetico
subprovince
ONTARIO
Thunder Bay
Isle Royale fault
SUPERIOR
Wawa subprovince
LAKE
SUPERIOR
Wawa Subprovince
Keweenaw fault
Duluth
Douglas fault
Syncline
Lake Superior
Malmo discontinuity
GLTZ
Benson block
Montevideo block
Morton block
Lake Owen fault
Niagara fault zone
Pembine-Wausau terrane
MICHIGAN
Eau Pleine shear zone
Marshfield terrane
GREEN BAY
WISCONSIN
SIMPLIFIED TECTONIC MAP OF PRECAMBRIAN
ROCKS, LAKE SUPERIOR REGION
0 30 60 KM.
MILES
Compiled by P.K.Sims, 1990

Figure 1

# EXPLANATION

## MAJOR PRECAMBRIAN TERRANES

| TECTONIC ELEMENT | | PRINCIPAL ROCK TYPES |
|---|---|---|
| Midcontinent rift system (1,100-1,050 Ma) | | |
| Late- and post-rift | Ys | Fluvial and lacustrine clastic sedimentary rocks |
| Syn-rift | Yv | Basalt, rhyolite, minor interflow sedimentary rocks, and gabbroic intrusion (Duluth Complex, Mellen Intrusive Complex) |
| Intrusive rocks of Transcontinental anorogenic province | | |
| Anorogenic intrusions | Yw | Wolf River batholith (1,470 Ma) |
| | Ygr | Granitic rocks |
| Anorogenic magmatism | Xrg | Rhyolite and cogenetic epizonal granite (~1,760 Ma) |
| Quartzite basins | Xq | Fluvial sand-dominated redbed sequences |
| Penokean orogen | | |
| Foredeeps | Xgs | Turbiditic graywacke-shale sequences |
| | Xpg | Passive-margin metasedimentary and metavolcanic rocks and overlying turbitic graywacke-shale |
| Fold and thrust belt | Xpm | Passive-margin metasedimentary and metavolcanic rocks, tectonically imbricated |
| | Xvg | Island-arc-related metavolcanic and granitoid rocks (1,890-1,840 Ma) and post-tectonic granitoid intrusions (~1,760 Ma) of Pembine-Wausau terrane |
| Magmatic terranes | XAi | Syn- to post-tectonic granitoid rocks intruded into complex metamorphic terrane |
| | XAr | Island-arc-related metavolcanic and granitoid rocks (~1,892-1,840 Ma) on Late Archean basement rocks |
| Superior craton | | |
| Greenstone-granite terranes (Superior province) | | |
| Wabigoon subprovince | Wvgg | Arc-related metavolcanic, metasedimentary, and syn-tectonic to late-tectonic granitoid rocks (2,750-2,650 Ma) |
| Quetico subprovince | Wsg | Turbidite-dominated metasedimentary rocks and granitoid intrusions (2,690-2,650 Ma) |
| Wawa subprovince | Wvg | Arc-related metavolcanic, metasedimentary, and syn-tectonic to late tectonic granitoid rocks (2,888-2,650 Ma) |
| Gneiss terrane | | |
| Supracrustal rocks | Wga | Interlayered bimodal metavolcanic rocks (2,750-2,640 Ma) |
| | Agn | Migmatitic gneiss and amphibolite of amphibolite to granulite metamorphic grade (3,600-2,700 Ma) intruded by late-tectonic to post-tectonic granite (~2,650 Ma) |

Note: Proterozoic Y (Middle Proterozoic) 1,600-800 Ma
Proterozoic X (Early Proterozoic) 2,500-1,600 Ma
Archean, A, 3,800-2,500 Ma; W, 2,800-2,500 Ma.

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# SUPERIOR PROVINCE GREENSTONE BELTS

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Concepts of greenstone belt geology have changed dramatically within the last few years. The former view was that Archean greenstone belts represented a single tectonic environment, either collapsed continental rifts (Goodwin 1981), back arc basins (Tarney et al. 1976), or amalgamated island arcs (Langford and Morin 1976). The predominant structural style was believed to be large scale synclinoria produced by the diapiric uprise of late granitic batholiths. The changes of the last few years are briefly reviewed below, concentrating on greenstone subprovinces of the Superior Province.

## LITHOSTRATIGRAPHIC ASSOCIATIONS

Four lithostratigraphic associations have been identified in greenstone belts of the Superior Province (Thurston and Chivers 1990):

1) Epicratonic platforms: Dated examples are >2.85 Ga and form at least three types (figure 1): a) epicratonic platform sequences with basal cross-bedded quartz arenites overlain by stromatolitic carbonates, succeeded by iron formation followed by komatiitic to tholeiitic volcanic units. b) quartz-rich submarine fan sequences of channelized conglomerates, sands and muds containing the cannibalized equivalents of the epicratonic platforms and grading basinward to less quartzose sands and muds, and c) komatiitic and tholeiitic flows with intercalated cross-bedded quartzose sands. The sequences are compared to modern passive margin and foredeep settings. All types occur in the greenstone belts of the Sachigo subprovince and the central part of the Wabigoon subprovince. Where undisturbed, they lie unconformably on older granitic rocks and greenstone sequences.

2) Mafic sequences: Generally <2.8 Ga, the mafic sequences consist of mafic and ultramafic flows with intercalated deep-water sulfidic argillites representing volcanism from a central volcanic complex.,The mafic sequences are compared to Archean oceanic, backarc, or the primitive stage of arc volcanism. These rocks rarely overlie older epicratonic platform sequences; more commonly they form the basal sequence of greenstone belts throughout all granite-greenstone subprovinces of the Superior Province (e.g. Kinojevis Group, Confederation Lake, Katiagamak Group.)

3) Arc volcanism: Mafic to felsic volcanic cycles (figure 2), <2.8 Ga, are the most common sequence type, forming major parts of greenstone belts of Abitibi, Wabigoon, Uchi, and Sachigo subprovinces. The volcanic style (bimodal, ash-flow dominated pyroclastic activity), petrographic, and geochemical evidence indicates these sequences represent Archean arc magmatism.

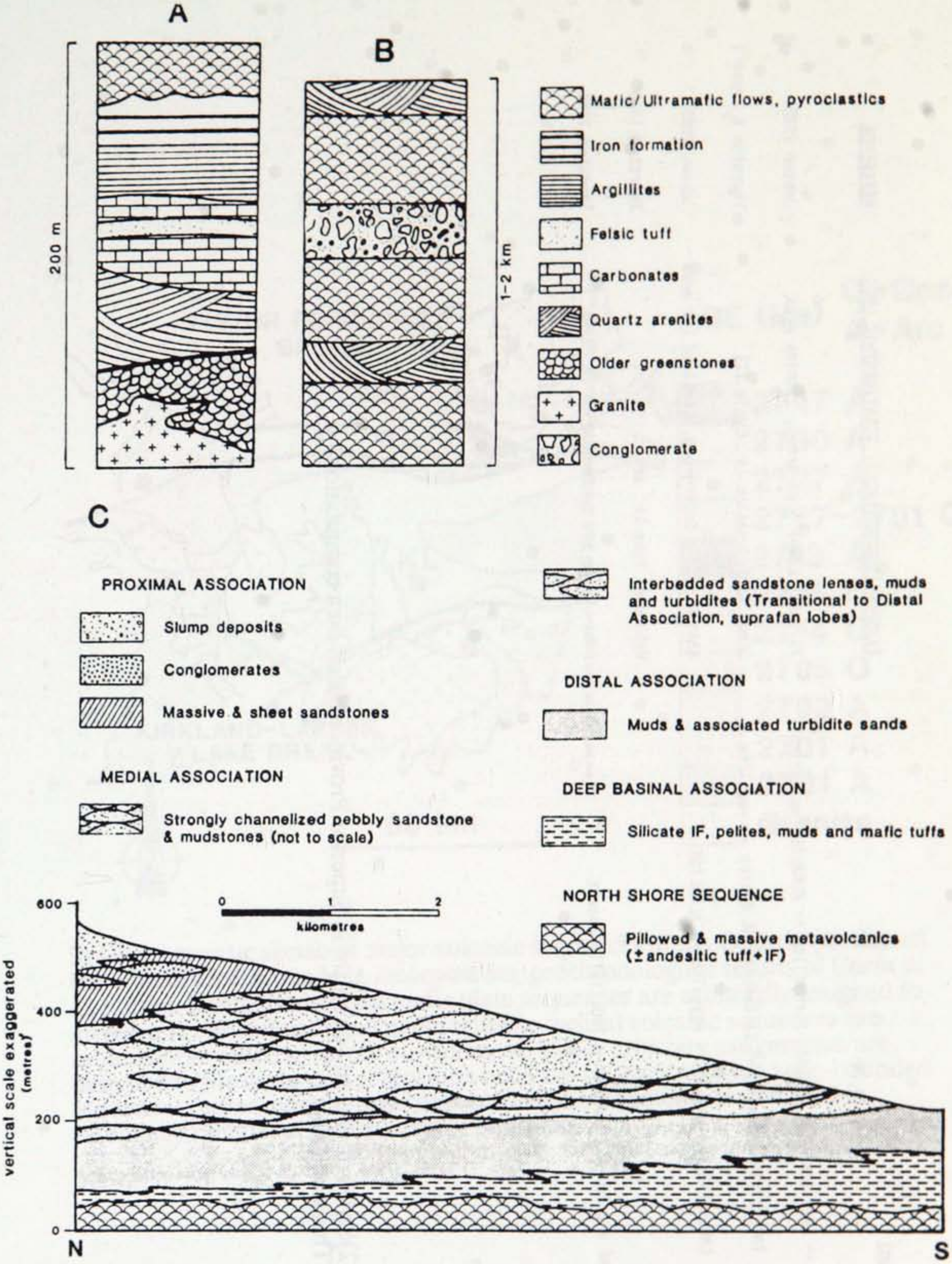
4) Late unconformable basins: These fault-bounded basins of alluvial-fluvial to deep marine metasediments and alkaline to calcalkaline, commonly subaerial metavolcanics are perhaps comparable to pull-apart basins or molasse units. They are generally <2.69 Ga in age. These sequences, where undisturbed, unconformably overlie older greenstones and are spatially associated with: a) major sequence boundaries within large greenstone belts (e.g. Timiskaming Group), or b) subprovince-bounding shear zones (e.g. Shebandowan Group).

## GREENSTONE BELT ARCHITECTURE

U/Pb age data and structural analysis have shown that greenstone belts consist of the above lithostratigraphic associations in either depositional or, more commonly, tectonic contact with one another. The common style is kilometer-scale homoclinal panels with uniform younging directions forming regional scale units. Greenstone belts display back-to-back and front-to-front relationships, separated by deformation zones. Age differences of several tens of millions of years (Confederation lake, Hemlo, Red Lake) and tectonic style contrasts between blocks suggest these relationships are the result of tectonic juxtaposition, and not due to folding. Out-of-sequence stratigraphy (e.g. Red Lake, Uchi Lake, Sioux Lookout, and Kirkland Lake) and large-scale homoclinal, fault-bounded blocks suggest that unrelated blocks were juxtaposed during the early stages of the deformation process. Greenstone belts are therefore viewed as tectonic collages representing a variety of environments, assembled prior to the development of the major shear zones separating granite-greenstone and sedimentary subprovinces.

## SYNTHESIS

The spatial and temporal distribution of the various lithostratigraphic associations represents a type of secular variation with early quartz-rich sequences, development of mafic sequences and arc sequences followed by "Late unconformable basins". The concentration of quartz-rich sequences and their great lateral extent in north-central Superior Province (figure 4) suggests the existance of a 2.9-3 Ga orogenic event in that region. Superior Province greenstone belts preserve evidence for 2.8 and 2.7 Ga orogenic events as well. The systematic southward younging of the age of volcanism, post-tectonic plutonism and shear zone development (Stott et al 1987) is consistent with the accretion of successively younger terranes onto the margin of a central ancient sialic nucleus to form the Superior Province as presently known.

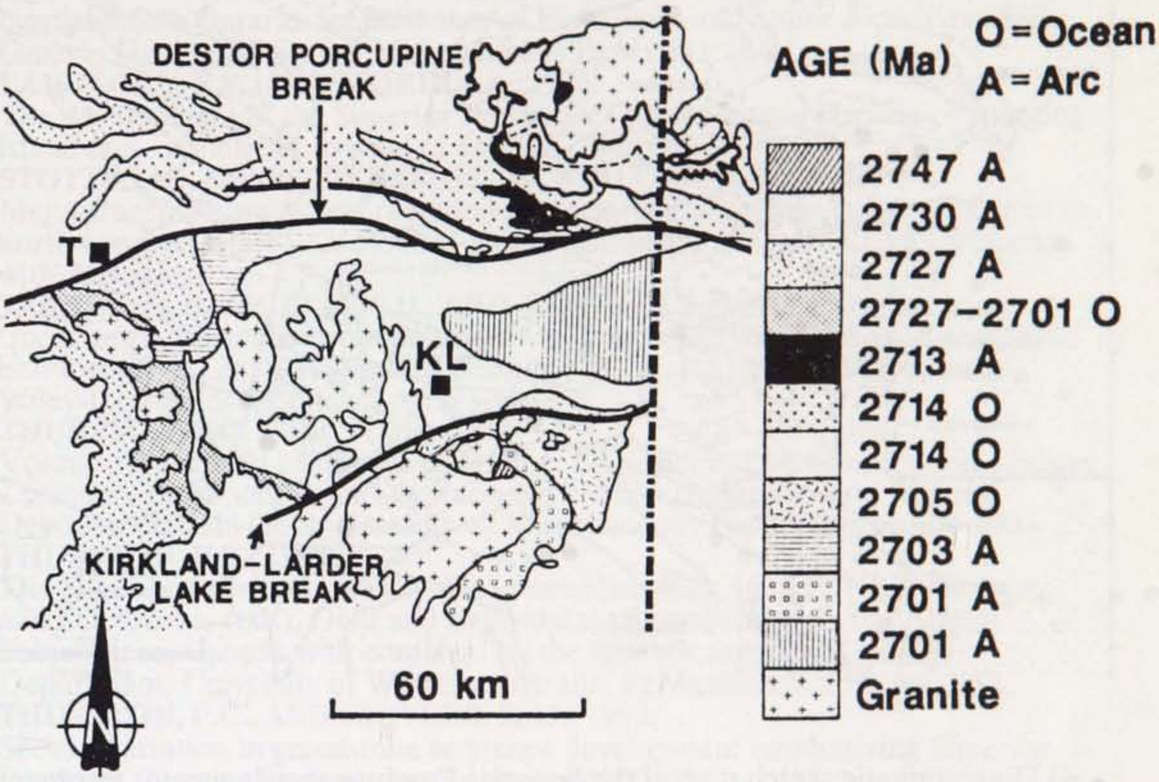



1) Diagrammatic stratigraphic sections of three types of quartz arenite-bearing greenstone lithostratigraphic associations. A: Undisturbed epicratonic platform sequence. B: volcanic associated platform sequence. C: Cross section of quartz-rich submarine fan composed in part of cannibalized platform sediments (after Cortis 1988). (Figure after Thurston 1990)

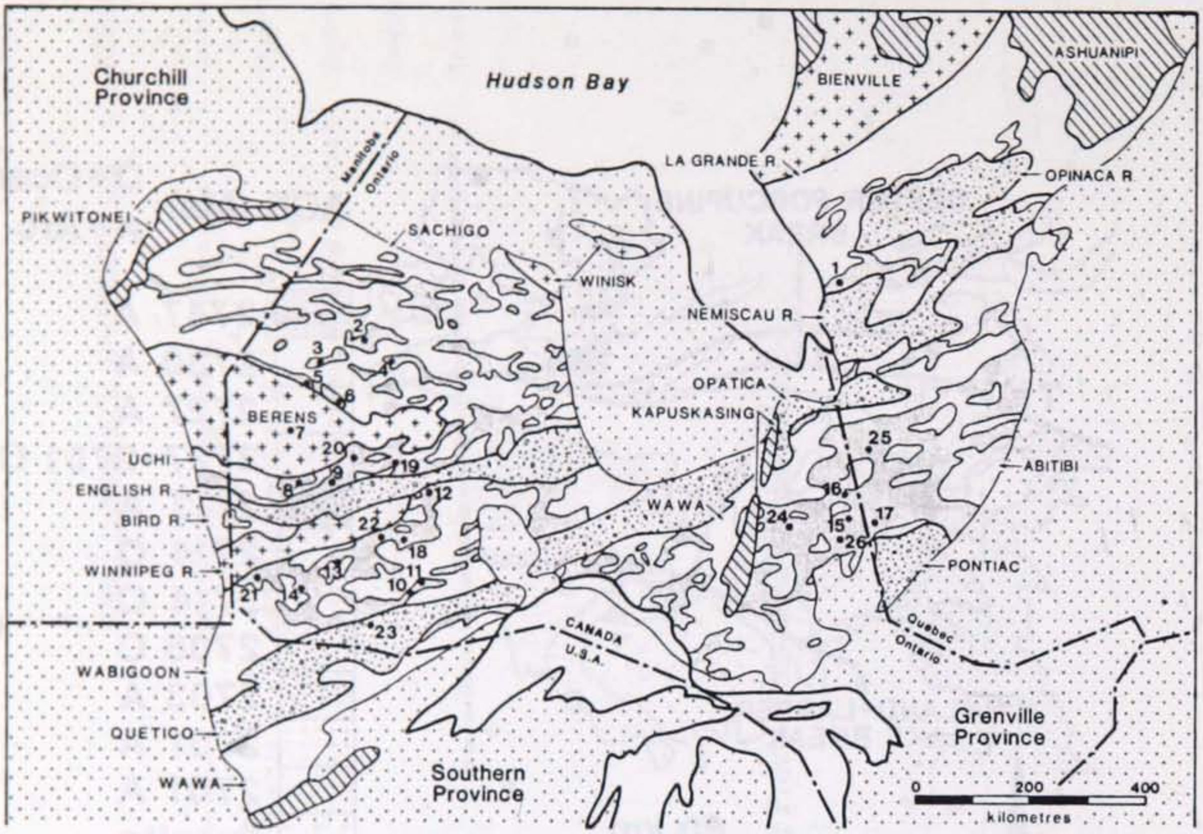
| TYPE (MAGMA CLANS are capitalized) | EXAMPLE (SUBPROVINCE): STRATIGRAPHIC UNIT | REFERENCE |
|---|---|---|
| - THOLEIITIC basalt -> rhyolite -- CALC-ALKALINE basalt -> rhyolite -- ALKALINE volcanics | Abitibi: Lower Supergroup, Timmins area | Jensen 1985 |
| THOLEIITIC basalt -> andesite -- THOLEIITIC andesite -- CALC-ALKALINE dacite -> rhyolite | Uchi: Confederation Lk. Cycle III | Thurston & Fryer 1 |
| THOLEIITIC basalt -> andesite -- CALC-ALKALINE basalt -> rhyolite -- ALKALINE | Abitibi: Uppr Supergroup, Kirkland Lk. area | Jensen 1985 |
| CALC-ALKALINE basalt -> rhyolite | Wabigoon: Yoke Lk. area | Thurston 1986 |
| THOLEIITIC basalt -- CALC-ALKALINE dacite -> rhyolite -- THOLEIITIC basalt | Uchi: Cycle III Confederation Lk. area | Thurston 1986 |

-> fractionation relationship
-- no fractionation relationship

2)Styles of cyclical volcanism in Superior Province based on a compilation by Thurston (1986).




3) Diagrammatic sketch of major volcanic sequences of the Ontario portion of the Abitibi greenstone belt, incorporating geochronological results of Corfu et al. (1989). Komatiite-bearing mafic plain sequences are arbitrarily assigned to an oceanic environment and mafic to felsic cyclical volcanic sequences are arbitrarily assigned to an arc environment. These arbitrary assignments are designed to illustrate that greenstone belts are collages of shear zone-bounded sequences representing diverse depositional environments asssembled in essentially random order by tectonic processes.




4) Diagrammatic sketch map of the Superior Province showing major terrane types. Plutonic terranes include: Berens Subprovince, Winnipeg River subprovince, and the Beinville subprovince. Sedimentary terranes include the English River subprovince, the Quetico Subprovince, and related Opatica and Nemiscau areas and the Pontian sediments. Granite-greenstone subprovinces include the Sachigo, Uchi, Bird River, Wabigoon and Abitibi subprovinces. Numbered localities are: *Platform sequences*: 1) Sakami Lake, 2) Muskrat Dam Lake, 3) Sandy Lake, 4) eyapamikama Lkae, 5) Favourable Lake, 6) North Spirit Lake, 7) McInnes Lake, 8) Red Lake, 9) Confederation Lake, 10) Steeprock Lake, 11) Lumby Lake, *Mafic sequences* 12) Jutten Group, 13) Atikwa Lake, 14) Katimiagamak Lake, 15) Stoughton-Roquemaure Group, 16) Kinojevis Grooup, 9) Confederation Lake area, *Arc sequences*: 17) Noranda, 18) Sturgeon Lake, 9) confederation Lake. *Late unconformable basins*: 19) Bamaji Lake, 20) Birch Lake, 21)Crow Duck 22) Ament Bay, 23) Seine 24) Timmins, 25 Casa Berardi, 26) Kirkland Lake.

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# ARCHEAN PLUTONIC ROCKS IN THE SOUTHERN SUPERIOR PROVINCE: MAGMATISM DURING ARC-CONSTRUCTION AND ARC-ACCRETION

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The southern Superior Province provides an unparalleled opportunity to examine large scale magmatic processes contributing to early crustal genesis. U-Pb geochronology (e.g. Davis *et al.* 1988; Corfu *et al.* 1989) and regional geological studies (e.g. Percival and Williams 1989) in the southern Superior Province have recently led to a model of accretion of island arcs, sedimentary prisms and continental slivers to explain the east-west trending metavolcanic-plutonic, metasedimentary and plutonic subprovinces (Williams 1990). This study examines the origin of plutonic suites within the southern Superior Province in the context of this model. Data reviewed for the preparation of the Geological Map of Ontario (Ontario Geological Survey 1990) are the basis of the interpretations here.

Granitoid rocks in the southern Superior Province, ranging in age from 3.17 to 2.65 Ga, are subdivided into 6 major suites: gneissic tonalite; foliated tonalite; two-mica granite (S-type); biotite granite-granodiorite; diorite-monzonite and diorite-syenite. Major field, petrographic and geochemical characteristics of these suites are summarized in Tables 1 and 2. Each of these six suites is associated with particular stages in the tectonic evolution of the accretionary process. **Foliated tonalite suite** plutons occur in several associations. Pre-tectonic 2.75-2.71 Ga tonalite plutons, encompassing a compositional range of diorite-tonalite-granodiorite and locally with layered gabbro cumulates, form composite batholiths spatially and temporally associated with calc-alkaline volcanism in volcanic-plutonic subprovinces. These rocks have geochemical attributes suggesting derivation from melting of a continuum between metasomatized mantle and basaltic slab in a subduction zone (Sutcliffe *et al.* 1990; Beakhouse and McNutt 1990). **Foliated tonalite suite** rocks also occur with **gneissic tonalite suite** rocks in 3.17-2.83 Ga complexes which are present in some metavolcanic-plutonic and plutonic subprovinces and predate the the main stage of volcanism (Davis et al. 1988). Both of these suites are also present as syn-tectonic 2.71-2.69 Ga complexes that were emplaced into mid-crustal levels (Percival and Krogh 1983) during the major thrust and fold deformation of metavolcanic-plutonic subprovinces (Jackson and Sutcliffe 1990). The syntectonic complexes form the voluminous high-silica sodic compositions with HREE depletion that are typically considered to be derived from partial melting of basaltic lithosphere (e.g. Martin 1986). The late- to post-tectonic 2.71-2.66 Ga **granite-granodiorite suite** forms large batholiths which are commonly microcline megacrystic and discordantly intrude tonalitic rocks in metavolcanic-plutonic and plutonic subprovinces. The elemental and isotopic composition of these rocks indicates that they are derived from melting of tectonically thickened assemblages of basalt and older tonalitic crust (Beakhouse and McNutt 1990). The late to post-tectonic 2.70-2.67 Ga **diorite-monzonite suite** is present in all subprovince types and includes a spectrum of silica-saturated to oversaturated compositions ranging from diorite to alkali granite with monzodiorite, monzonite and granodiorite predominating (Stern *et al.* 1989). These rocks have compositional similarities to pre-tectonic calc-alkaline tonalite suite rocks but are distinquished by higher abundances of LILE and range from calc-alkaline to shoshonitic. The suite is associated with comagmatic lamprophyre dikes, amphibole-rich cumulate and dike rocks (appinites) and locally ultramafic to gabbroic intrusions (Sutcliffe et al. 1990). Elemental and isotopic characteristics indicate the diorite-monzonite suite and associated rocks are derived from enriched mantle sources (Shirey and Hanson 1984) and have compositions modified by amphibole fractionation (Sutcliffe et al. 1990) and crustal assimilation (Gariepy and Allegre 1985). **Two-mica granite suite** rocks with "S-type" geochemical attributes are largely restricted to the metasedimentary subprovinces that are interpreted to be accretionary prisms which developed adjacent to coeval volcanic arcs. The late- to post-tectonic two-mica granites, ranging in age from 2.67-2.65 Ga, are derived from from melting of metasedimentary sources (Sawyer 1987; Percival and Williams 1989; Breaks 1989). The post-tectonic (<2.68 Ga) silica-saturated to undersaturated **diorite-syenite suite** is compositionally gradational with the LILE-enriched diorite-monzonite suite, with which it represents a continuum of melting of progressively less hydrous and more LILE-enriched mantle sources (Sutcliffe *et al.* 1990).

The evolution of mantle- and crustal-derived melts in the Archean terranes of the southern Superior Province can largely be viewed as analogous to processes in Phanerozoic arc development and arc accretion. Foliated tonalite and gneissic tonalite suite plutons older than 2.8 Ga represent cratonic slivers that predate the main arc development. Foliated tonalite suite rocks that occur as composite plutons associated with volcanic rocks formed during the main period of arc construction and predate tectonism. Syntectonic 2.7 Ga gneissic and foliated tonalite suite rocks were emplaced during collision and with thrusting, contributed to crustal thickening. Late- to post-tectonic, mantle-derived diorite-monzonite and diorite-syenite suite plutons and associated shoshonitic to alkaline volcanic rocks are similar to those developed in the terminal stages of subduction in Phanerozoic regimes. These mantle-derived suites are coeval with late- to post-tectonic granite-granodiorite and two-mica granite suite rocks that postdate arc-arc and arc-continent collision and are derived from melting of tectonically thickened crust.

**Acknowledgement**
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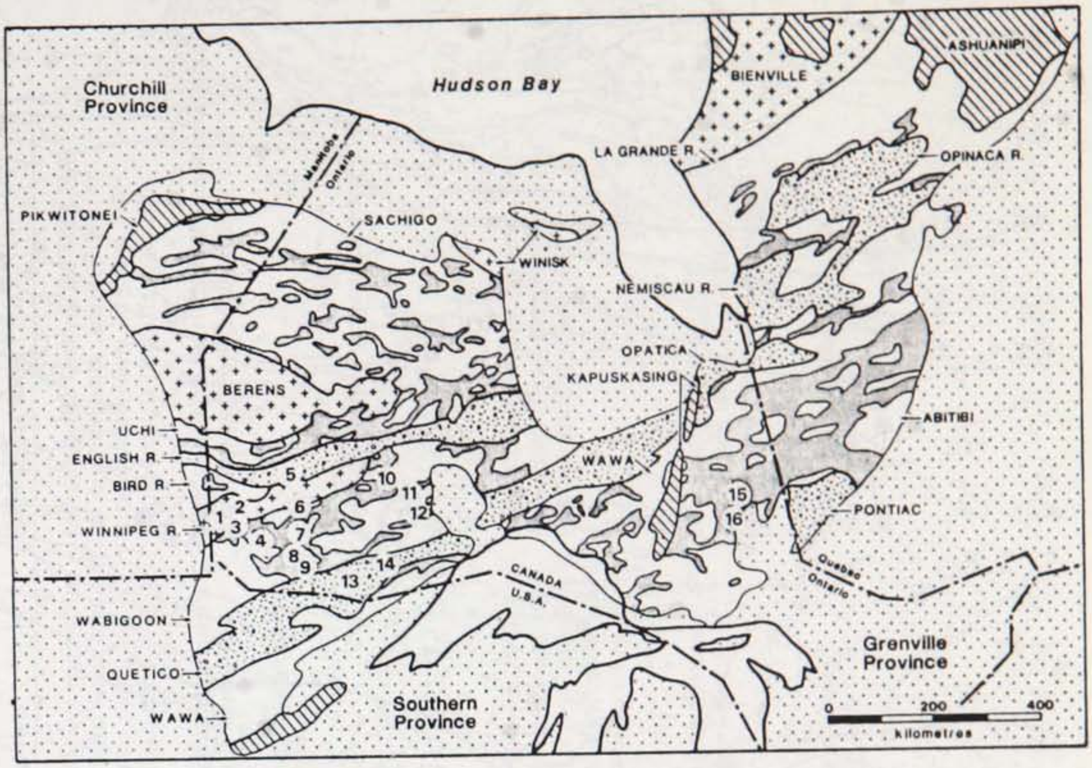


Figure 1. Subprovinces of the Superior Province (Card and Ciesielski 1986) showing locations of representative granitoid suites named in the tables. Foliated tonalite suites: Aulneau-4; Atikwa batholith-7; Lac des Iles tonalites-12; Abitibi batholith-15. Gneissic tonalite suites: Kenora area-3; Rainy Lake-9. Granite-granodiorite suite: Lount Lake batholith-2. Diorite-monzonite suite: Trout Lake pluton-1; Jackfish Lake pluton-8; Roaring River complex-11. Two-mica granite suite: Perrault Falls-5; Ghost Lake pluton-6; Quetico migmatites-14. Diorite-syenite suite; Sturgeon Narrows-10; Poohpah Lake-13; Otto stock-16.

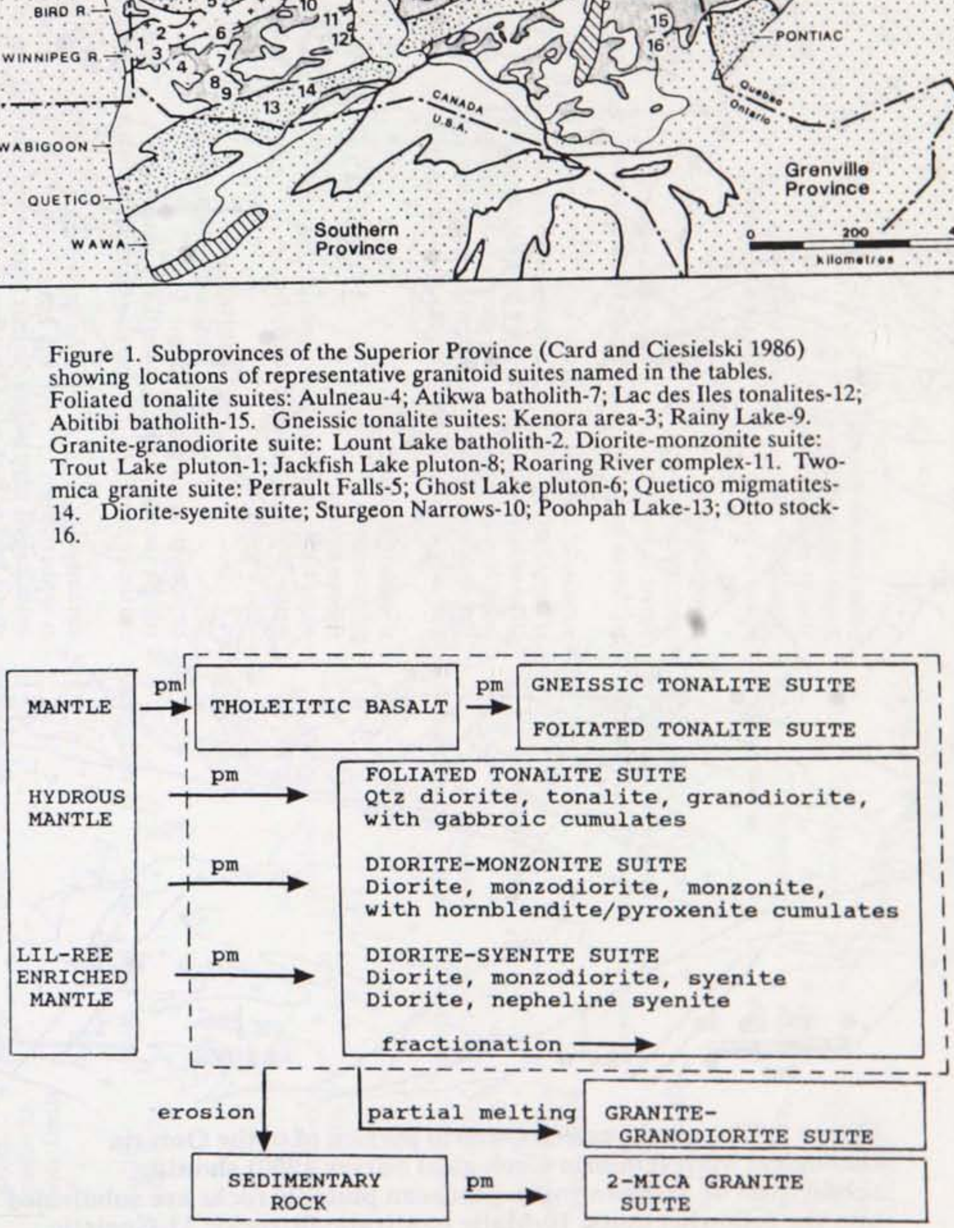


Figure 2. Schematic petrogenetic processes in the formation of Late Archean granitoid suites of the southern Superior Province.

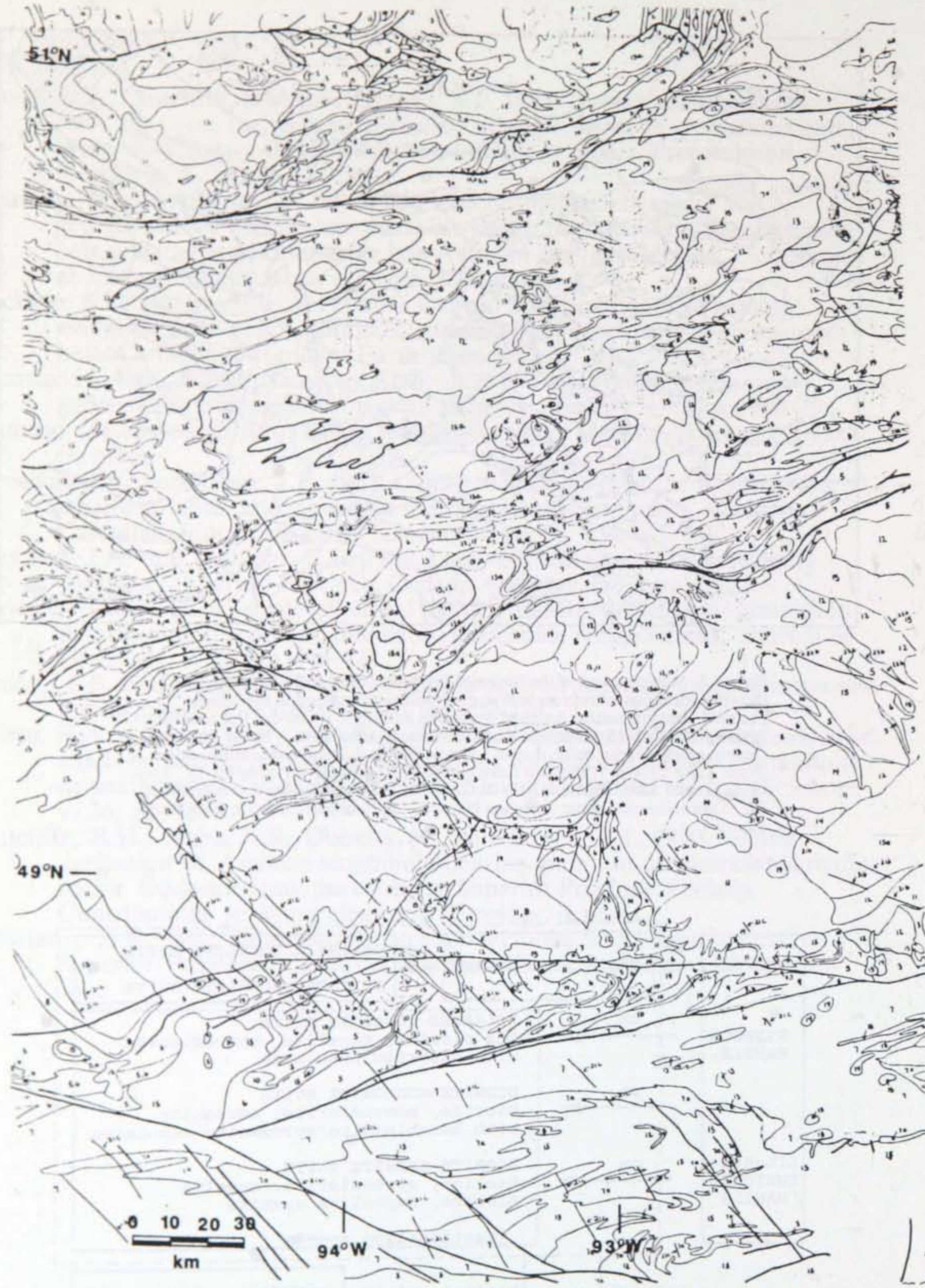


**Figure 3. The northwestern Ontario portion of of the Ontario Geological Map (Ontario Geological Survey 1990) showing subdivision of Archean rocks. Archean plutonic rocks are subdivided into the following units: 10-Mafic to ultramafic rocks; 11-Gneissic tonalite suite; 12-Foliated tonalite suite; 13-Two mica-granite suite; 14-Diorite-monzonite suite; 15-Granite-granodiorite suite; 16-Diorite-syenite suite. Units 1 to 9 are Archean supracrustal rocks. Units 17 to 23 are Proterozoic rocks. Area shown is bounded by latitudes 48o to 51oN and 92o to 95oW.**

Table 1: Field and petrographic characteristics of granitoid plutonic suites in the southern Superior Province

| Suite | Rock types* | Mafic minerals | Feldspars | Accessories | Textures | Associated rocks | Representative Suites |
|---|---|---|---|---|---|---|---|
| Foliated tonalite | tonalite, granodiorite, quartz diorite | bio±hb | pl±mc | sph,zir,ap,mgt | usually foliated, med. to coarse grained, hypidiomorphic or recrystallized; equigranular | often coeval with calc-alkaline volcanic rocks; locally associated with gabbroic plutons; magma mixing textures | Aulneau Batholith (Beakhouse and McNutt 1990)<br>Lac des Iles tonalite (Sutcliffe 1989)<br>Atikwa Batholith (Davis and Edwards 1985)<br>Abitibi Batholith (Sutcliffe et al. 1990) |
| Gneissic tonalite | tonalite, granodiorite | bio±hb | pl±mc | sph,zir,mgt | gneissic to strongly foliated, recrystallized; equigranular, medium grained | amphibolite/diorite enclaves; locally with deformed mafic dikes; leuco-tonalite and pegmatitic mobilizates | Rainy Lake tonalite (Sutcliffe 1980; Schwerdtner 1990)<br>Kenora area gneisses (Gower et al. 1982) |
| Granite -granodiorite | granite, granodiorite, tonalite | bio | pl,mc (subsolvus) | hb,mgt,sph, zir,ap | massive to foliated (primary), medium grained to pegmatitic; hypidiomorphic to allotriomorphic; mc commonly megacrystic | pegmatites, tonalitic cumulates | Lount Lake Batholith (Beakhouse and McNutt 1990) |
| Diorite -monzonite | monzodiorite, granodiorite, monzonite, diorite, quartz diorite, alkali granite | hb±bio±aug rarely opx | pl,mc (subsolvus); per (hypersolvus) | ap,sph,mgt,zir | massive to foliated (primary) locally layered, medium to coarse grained, hypidiomorphic, mc interstitial to megacrystic, clotty mafics | hornblendite/pyroxenite enclaves; amphibole-rich (appinite) dikes and cumulate rocks; lamprophyre dikes; locally ultramafic-gabbroic intrusions; magma mixing textures | Jackfish Lake Pluton (Sutcliffe et al. 1990)<br>Roaring River Complex (Stern et al. 1989)<br>Trout Lake Pluton (Beakhouse 1983) |
| Two-mica granite | granite, granodiorite, lesser tonalite, monzonite | bio±musc | pl,mc (subsolvus) | gt,cord,sill tour, mgt, ap | foliated to massive, med grained to pegmatitic allotriomorphic, inequigranular | metasedimentary schlieren; rare element pegmatites | Ghost Lake (Breaks et al. 1985)<br>Perrault Lake (Breaks et al. 1985)<br>Quetico Subprovince migmatites (Sawyer 1987) |
| Diorite -syenite | syenite, monzodiorite, diorite, neph. syenite, mela-syenite, malignite | aug±amph ±bio ±mel gt | per (hypersolvus); pl±mc (subsolvus) ±ne; rarely or | ap,sph,mgt,zir | massive to foliated (trachytoid), locally layered, medium to coarse grained, hypidiomorphic, K-spar often megacrystic | pyroxenite enclaves; lamprophyre dikes | Sturgeon Narrows Complex (Sage 1988)<br>Poohbah Complex (Sage 1988)<br>Otto Stock (Sutcliffe et al. 1990) |

Abbreviations: bio=biotite; hb=hornblende; musc=muscovite; aug=augite; opx=orthopyroxene; amph=amphibole; gt=garnet (mel-melanitic);
pl=plagioclase; mc=microcline; per=perthite; or=orthoclase; ne=nepheline; sph=sphene; zir=zircon; mgt=magnetite;
cord=cordierite; sill=sillimanite; tour=tourmaline

*Rock types are listed in approximate order of decreasing abundance.

Table 2. Summary of geochemical characteristics of representative sites of granitoid rocks in the Southern Superior Province

| Suite | Foliated tonalite | | Gneissic tonalite | Granite-granodiorite | Diorite-monzonite | Two Mica Granite | Diorite-syenite |
|---|---|---|---|---|---|---|---|
| | Aulneau | Lac des Iles | Rainy Lake | Lount Lake | Jackfish Lake Pluton | Ghost Lake | Otto Stock |
| $SiO_2$ wt% | 70-72 | 63-72 | 70-74 | 64-73 | 54-70 | 70-76 | 64-70 |
| mol $Al_2O_3$/ ($CaO+Na_2O+K_2O$) | ~1 | ~1 | ~1 | ~1 | generally<1 | generally>1 | generally<1 |
| mol $Na_2O/K_2O$ | high 3-10 | high 4-8 | mod 3-6 | low-mod 0.5-2 | mod/high 1.5-8 | low-mod 2-0.2 | mod ~2 |
| Sr ppm | 420-620 | 250-350 | 340-870 | 50-750 | 250-1540 | 25-300 | 340-380 |
| Rb ppm | <100 | <50 | <65 | 50-200 | <100 | 100-500 | 60-110 |
| $Ce_N$ | 30-60 | 35-70 | 20-40 | 30-200 | 60-180 | 60-80 | 20-30 |
| $(Ce/Yb)_N$ | 21-35 | 8-17 | 20-26 | 3-60 | 20-30 | 3-60 | 9-11 |
| Eu anomaly | negligible | negligible | negligible | -ve | none | -ve | none |
| Rad. Isotopes | mantle | n/a | n/a | crustal | mantle [6] | n/a | mantle |
| Other | -- | -- | HREE depletion | -- | high Cr,Ni with low $SiO_2$ | B,Be,Ga,Nb Li,Sn enrichment | locally $SiO_2$ undersat |
| Age Ga | 2.72-2.71 | 2.73 [5] | >2.70 | 2.70 | 2.70 [5] | 2.68 | 2.68 |
| References | (1) | (2) | (3) | (1) | (2) | (4) | (2) |

1-Beakhouse and McNutt (1990); 2-Sutcliffe et al (1990); 3-Sutcliffe (1980-unpublished thesis); 4-Breaks (1989-unpublished thesis); 5-D.W. Davis (unpublished data); 6-Shirey and Hanson (1984)

# STRUCTURES IN ARCHEAN ROCKS OF THE SOUTHERN SUPERIOR PROVINCE

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The Archean rocks of the Superior Province (3,000 - 2,650 Ma) of the Canadian Shield display structures similar to those found in ancient cratons around the world. There are two major types of subprovince (Fig. 1), high-grade metasedimentary gneiss belts and low-grade volcanic-plutonic belts (including greenstone belts). I will describe the characteristic structures found in these rocks, with particular emphasis on greenstone belts, and discuss their interpretation in terms of tectonic processes. Most of the examples are taken from the two margins of the Vermilion granitic complex (part of the Quetico gneiss belt) in northern Minnesota. On the southern side is the Vermilion district, a granite-greenstone belt that is part of the Wawa subprovince (Figs. 1, 2). On the northern side is the Rainy Lake wrench zone and the Wabigoon subprovince (Fig. 12).

In the the southern Superior Province, there is usually a *dominant cleavage or schistosity* (Fig. 3) in the volcanic and sedimentary rocks of the greenstone belts. This is fairly constant in trend, striking parallel to the boundaries of the belt, and steeply dipping. This cleavage is axial planar to folds in bedding ($S_0$) on all scales. The folds are typically asymmetrical, open to tight, and cleavage and bedding are nearly parallel on the long limbs of the folds. A mineral lineation is sub-parallel to the folds hinges, which plunge moderately to steeply in both directions in the plane of the cleavage (Fig. 3). Clasts in the sedimentary and volcanic rocks are strained, being flattened in the cleavage and stretched parallel to the mineral lineation. There is often variation in the shape of the strain ellipsoid from oblate (pancake) shape to prolate (cigar) shape, depending on position within the belt, especially with respect to major boundaries, or position with respect to major structures (Fig. 4).

Structures predating the dominant cleavage and associated folds are often subtle, and are best revealed by a careful study of *structural facing* (Fig. 5) in the dominant cleavage. Large areas of inverted strata have been found (Poulsen et al., 1980) and early isoclinal folds ($F_1$) have been revealed by patterns of reversals in structural facing (Fig. 2). A remarkable feature about these early folds is that there is often no cleavage ($S_1$) associated with them. At the time of fold formation the sediments were presumably unlithified or only partially lithified. They were also presumably recumbent. Because they are often large structures they may have been nappes (Bauer, 1985). Thrusts and repeated stratigraphy have been described in a greenstone belt in South Africa (de Wit, 1982).

Because of the presence of early folds, the dominant cleavage formed in a second deformational event ($D_2$). It is thus usually designated $S_2$ and the associated folds $F_2$. It is common to find $S_2$ and $F_2$ deformed by younger folds. $S_2$ may become crenulated to form $S_2'$ or $S_3$.

Correlation of the dominant cleavage can often be made across subprovince boundaries, but it is uncertain to what extent early structures can be correlated across across these boundaries. In the high-grade gneiss terranes, the structural picture is more complicated. Bedding and cleavage are more variable in attitude. Additional phases of folding can be identified, both earlier and later than $F_2$, and these are often spatially associated with granitoid intrusions.

There are two main processes that have been proposed to account for the structures in greenstone belts, *diapirism* and, more recently, *transpression*. Concentric patterns of foliation in large plutons in many shields suggest emplacement by diapirism (Fig. 6). Foliation typically dips more steeply with increasing distance from the centers of the plutons, and the concentric pattern of foliation may be matched by one of strained xenoliths (Fig. 7). A greenstone belt between two large diapirs may display rotated bedding and strong vertical stretching (Fig. 8). The pattern of structures displayed will depend on the level of exposure (Fig. 8). Shear zones of dip-slip sense (Fig. 8) along the margins of the plutons will be induced by diapirism. Alternatively, inflation of the pluton by ballooning (Ramsay, 1980) will involve no shear along the margin.

There is good evidence for shear (Fig. 9) in the rocks of the southern Superior Province, but this shear is *strike-slip* in sense, not dip-slip. In fact, both boundaries of the Quetico gneiss belt have experienced strong dextral shear in association with north-south shortening -- in other words a dextral transpression. Transpression can account for the pattern of oblate and prolate strains (Fig. 10), the single sense of asymmetry of folds associated with the dominant cleavage (Fig. 2), the local development of secondary cleavage parallel to the dominant cleavage (Fig. 11), and the presence of sigmoidal patterns of foliation and faulting on small and large scales (Fig. 12).

The main structural imprint in the rocks at the boundaries of the Quetico belt resulted from a dextral transpression. Many mapped and unmapped faults and shear zones were active during this event. Pluton emplacement played a secondary role in structure and fabric development. For transpression to have occurred, these boundaries would have had to have been regions of relatively soft lithosphere.

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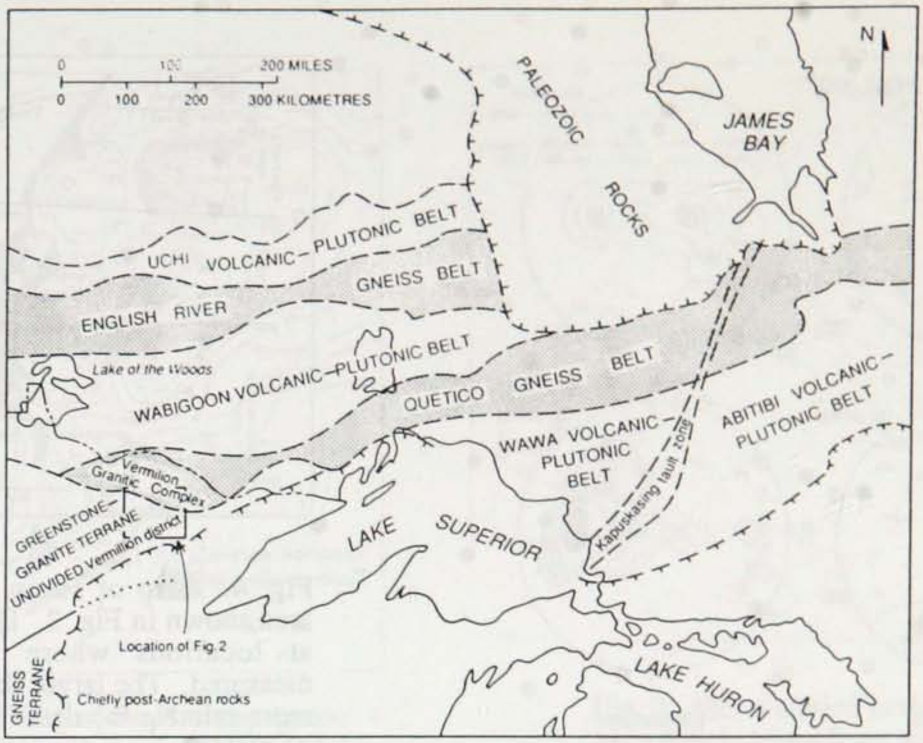


Fig. 1. Map of the Lake Superior region showing subprovinces of the Superior Province and the location of the Vermilion district (From Hudleston *et al.*, 1988).

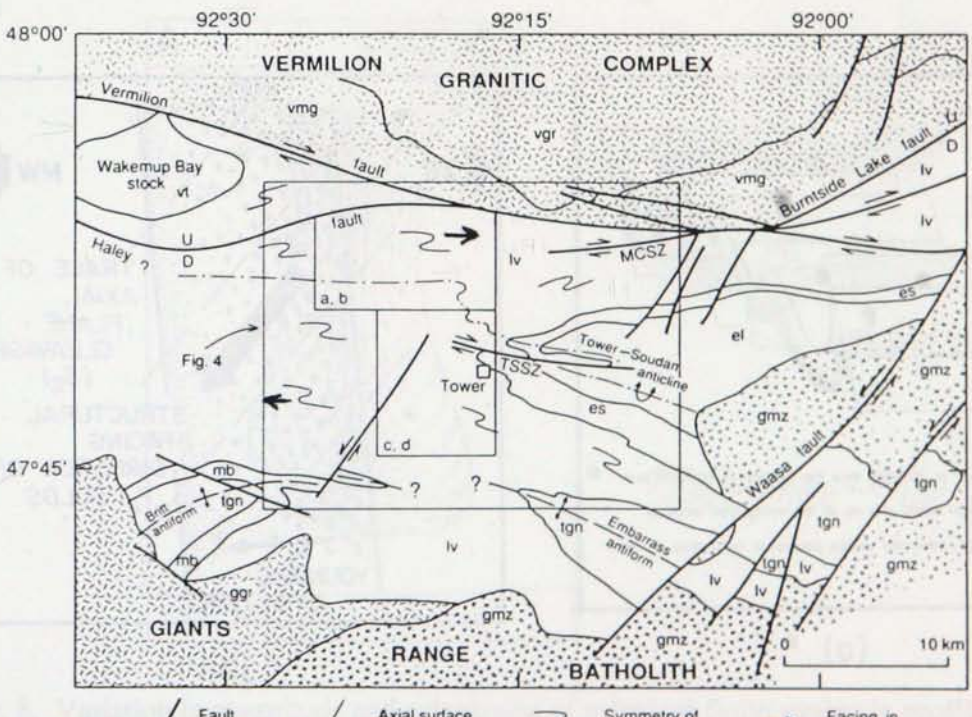


FIG. 2. Simplified geological map of the Vermilion district. Rock units: vgr, granite, chiefly the Lac La Croix; vmg, biotite schist, paragneiss, and migmatite; vt, tonalite; lv, Lake Vermilion Formation plus closely associated rocks of the upper member of the Ely Greenstone; es, Soudan Iron Formation (a member of the Ely Greenstone); el = Lower Member of the Ely Greenstone; mb = metabasalt; tgn = tonalite gneiss; ggr = granite; gmz = granite and monzonite; MCSZ = Mud Creek Shear Zone; TSSZ = Tower−Soudan Shear Zone.

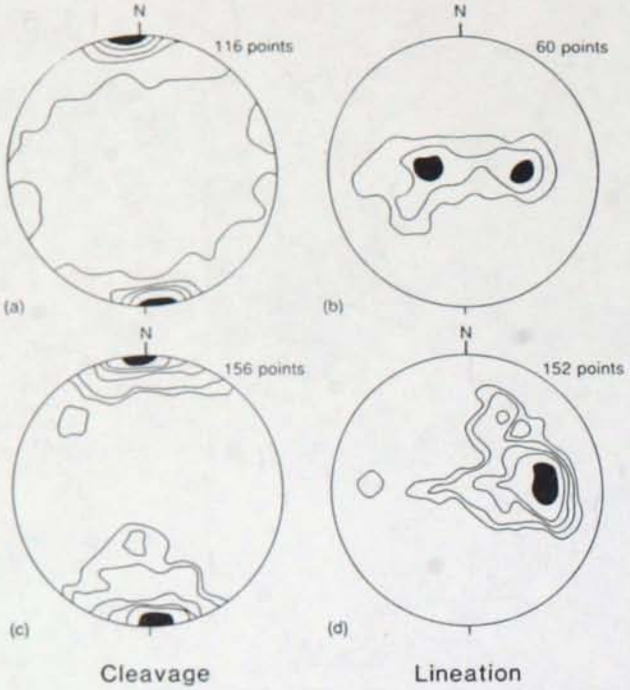


Fig. 3. Structural data for the two areas outlined in Fig. 2. (*a,c*) Poles to $S_2$ foliation; (*b,d*) $L_2$ lineations, including mineral lineation, $S_2/S_0$ intersections, and fold hinges (From Hudleston *et al.*, 1988).

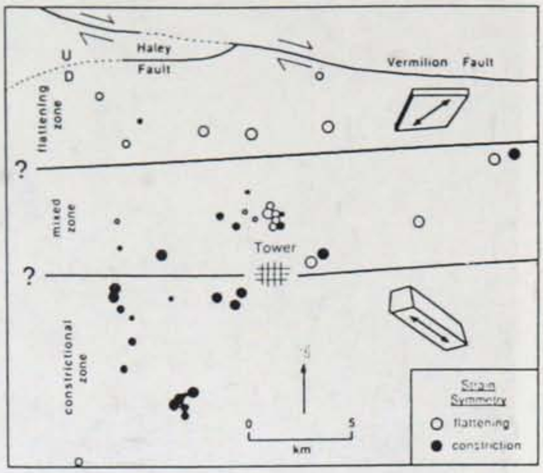


Fig. 4. Map of strain symmetry for the area shown in Fig. 2. Dots and circles are at locations where strain has been measured. The larger the dot or circle, the more reliable the data (From Hudleston *et al.*, 1988).

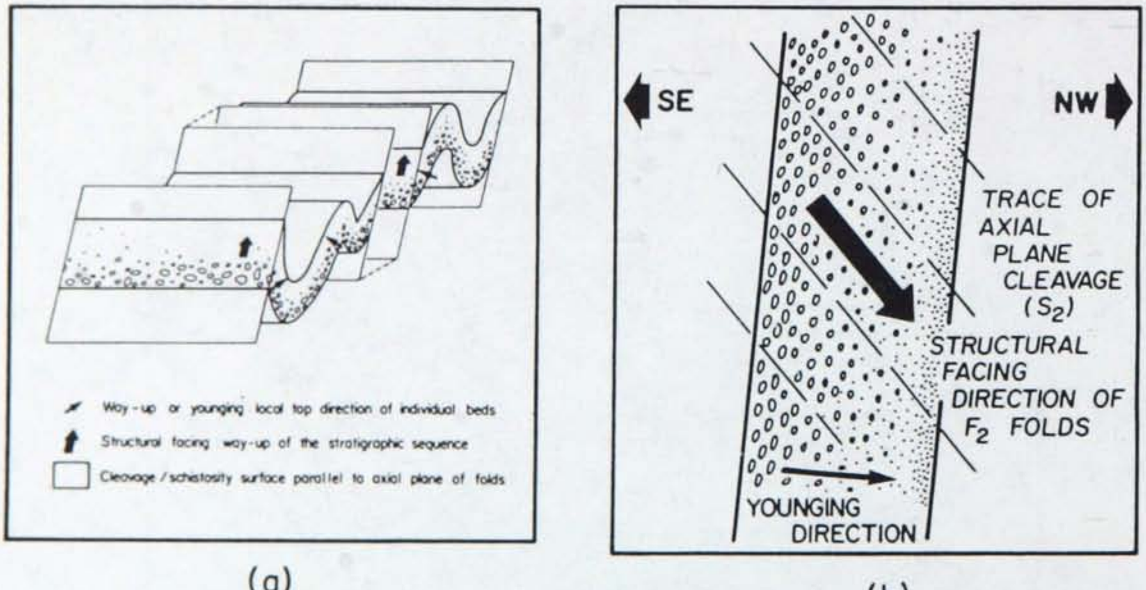


Fig. 5. The structural facing of folds. (*a*) A general illustration of the concept of structural facing. The structural facing direction remains consistently oriented while younging directions are variable. (*b*) A schematic profile view of bedding and cleavage. $F_2$ folds face downward since younger beds are encountered downward along the $S_2$ cleavage (After Poulsen *et al.*, 1980).

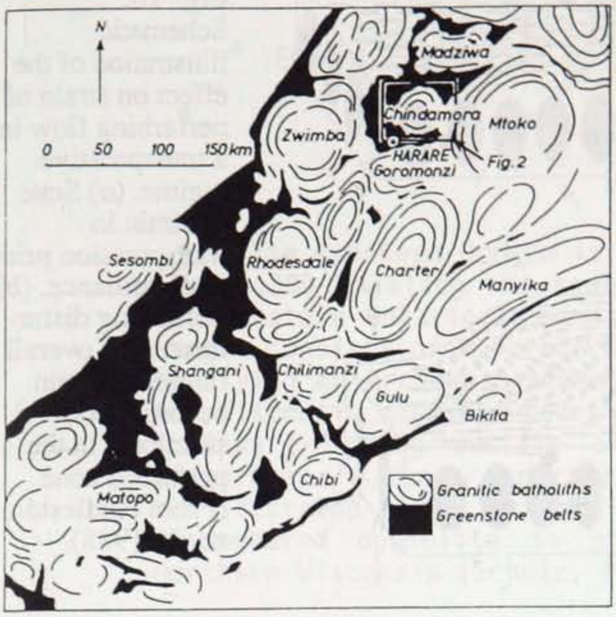


Fig. 6. General map of the Archean craton of Zimbabwe showing the distribution of granitic batholiths and greenstone belts (From Ramsay, 1989).

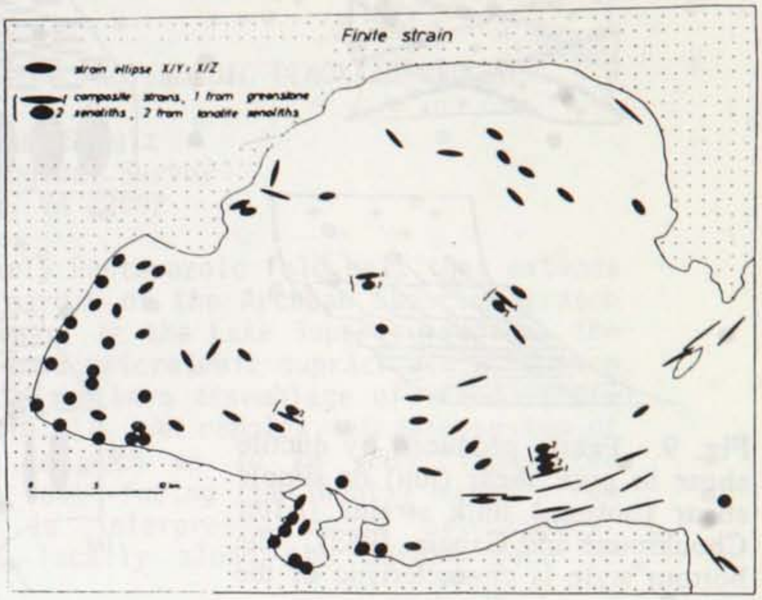


Fig. 7. Finite strains in the Chindamora batholith deduced from an analysis of xenolith shapes (From Ramsay, 1980).

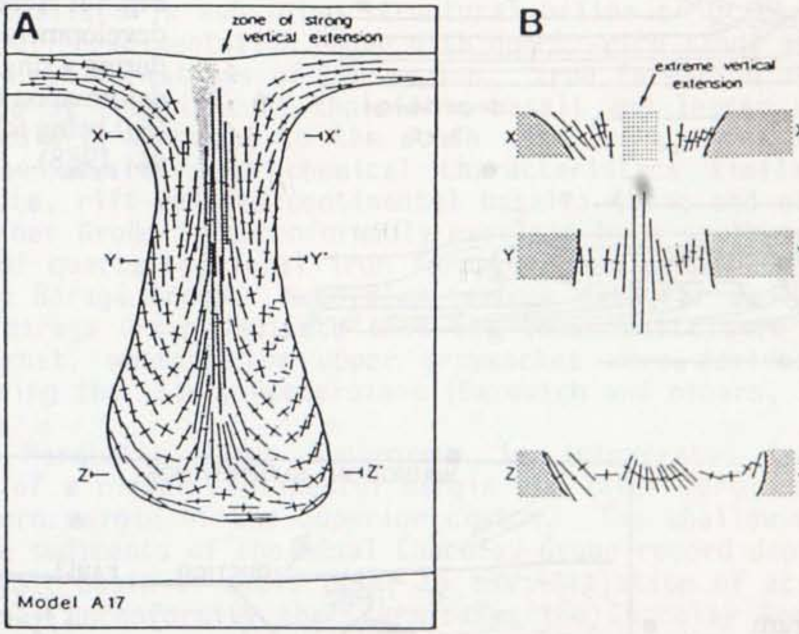


Fig. 8. Variation in magnitude and orientation of principal finite strains in profile section of a mature subsided trough. (*a*) Strain variation throughout the trough. (*b*) Strain variation across three levels of the trough indicated in (*a*) (From physical model of Dixon and Summers, 1983).

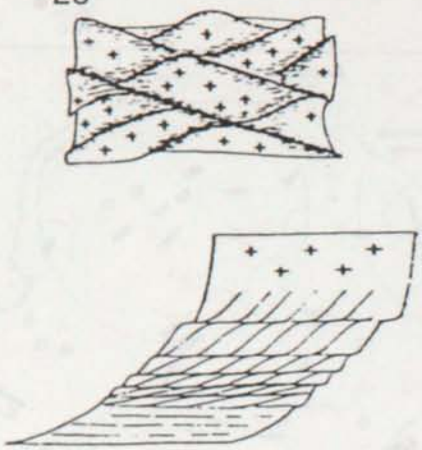

Fig. 9. Fabric produced by ductile shear in pure shear (top) or simple shear (bottom) bulk strains (After Choukroune and Gapais, 1983). The bottom style is characteristic of the southern Superior Province.

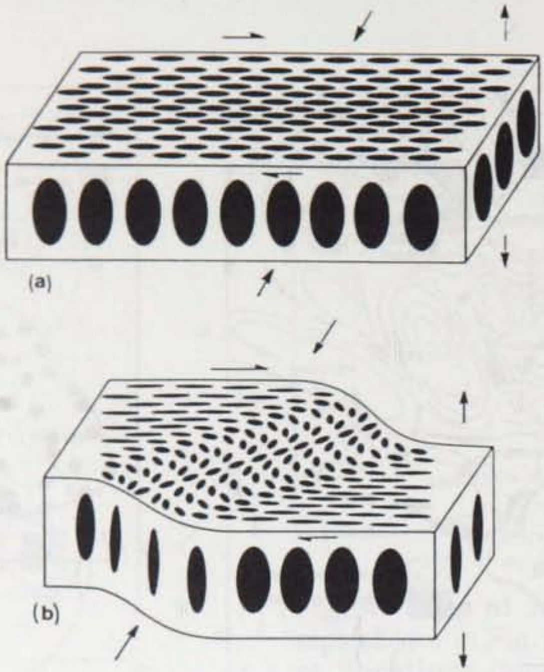


Fig. 10. Schematic illustration of the effect on strain of perturbing flow in a transpressive regime. (*a*) State of strain in transpression prior to disturbance. (*b*) Following disturbance, the overall flattening strain becomes constrictional in the perturbed zone (From Hudleston *et al.*, 1988).

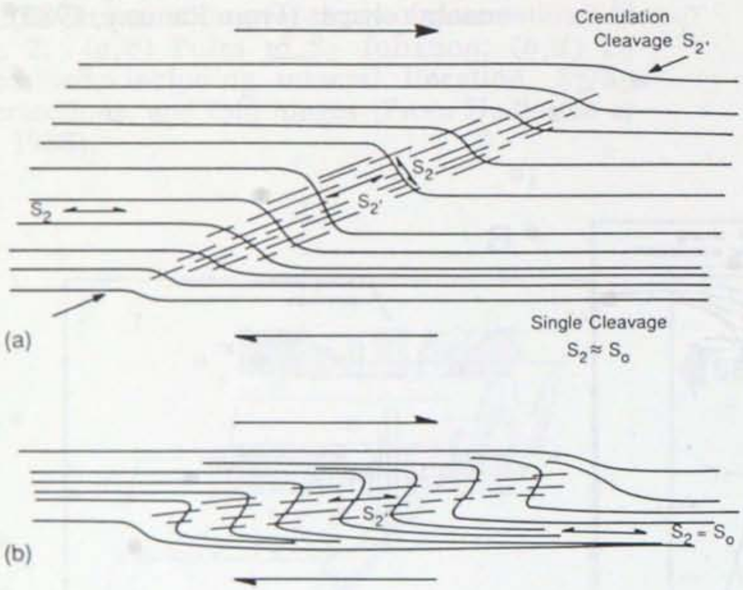


Fig. 11. Schematic illustration of the development of folds and $S_2'$ cleavage during a single transpressive deformation that produced the foliation ($S_2$) being folded (From Hudleston *et al.*, 1988).

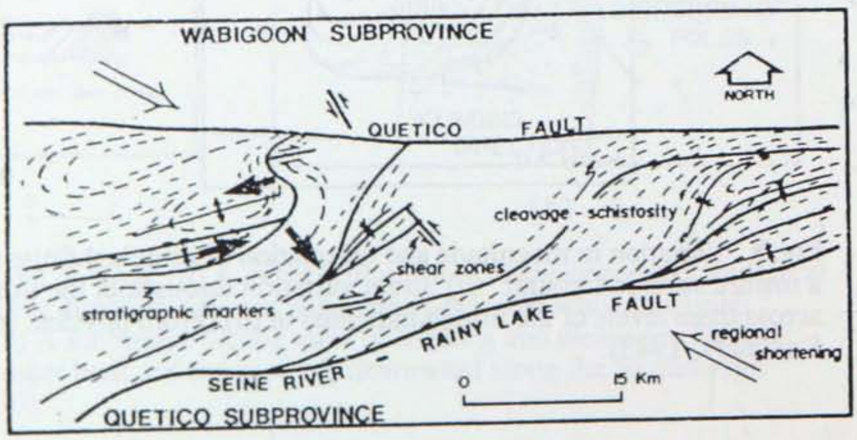


Fig. 12. Schematic diagram illustrating structural features of the Rainy Lake Wrench Zone. Short solid arrows identify downward facing units (After Poulsen, 1986).

# TECTONIC EVOLUTION OF THE EARLY PROTEROZOIC PENOKEAN OROGEN

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The Penokean Orogen is an Early Proterozoic fold belt that extends some 1,300 km along the southern margin of the Archean Superior craton from eastern Ontario to western Iowa. In the Lake Superior region, the orogen consists of a northern deformed epicratonic supracrustal sequence overlying an Archean basement and a southern assemblage of island arcs, the Wisconsin magmatic terranes (Fig. 1). An complex, arcuate system of southward-dipping ductile shears as much as 10 km wide, named the Niagara fault zone, separates the south-facing continental margin from the accreted arc terranes and is interpreted as a paleosuture. Dismembered ophiolite is present locally along the suture zone in northern Wisconsin (Schulz, 1987).

In Michigan, the Early Proterozoic epicratonic sequence, the Marquette Range Supergroup, consists mainly of clastic sedimentary rocks and tholeiitic basalt with subordinate chemical sediments, chiefly iron formation. These units were deposited during the period from about 2100 to 1850 Ma. The basal Chocolay Group consists of shallow- water quartzite and dolomite (Larue, 1981) with stable shelf characteristics. Deposition of these units was followed by a period of variable uplift and erosion that resulted in a regional angular unconformity and a marine transgression with sedimentation, represented by the Menominee Group, localized in subsiding structural basins or grabens (Larue and Sloss, 1980). Sedimentation began with quartz-rich sands and ended with the major iron formations of the region. Iron formation deposition was accompanied by eruptions of tholeiitic basalt and lesser rhyolite that are particularly abundant to the south near the present Niagara fault zone. The basalts have chemical characteristics similar to recent within-plate, rift-related continental basalts (Ueng and others, 1988). The Menominee Group is unconformably overlain by a southward thickening sequence of quartzite, local iron formation and basalt, and graywacke-slate, the Baraga Group. Neodymium isotope data for sedimentary rocks from the Baraga Group indicate that the lower units were derived from Archean crust, whereas the upper graywackes were derived from crust formed during the Early Proterozoic (Barovich and others, 1989).

The Marquette Range Supergroup is interpreted to record the evolution of a rifted continental margin and later foreland basin along the southern margin of the Superior craton. The shallow-water, stable shelf-type sediments of the basal Chocolay Group record deposition in an intracratonic basin or shelf prior to the initiation of active rifting. The regional unconformity that terminates the Chocolay Group marks the begining of major extension and is equated with the rift-onset unconformity characteristic of recent rifted continental margins (Falvey, 1974). The succeeding Menominee Group, deposited in restricted, rapidly deepening grabens, represents a period of active rifting that preceded actual crustal separation and sea floor spreading. The unconformity at the top of the Menominee Group is equated with the breakup unconformity of recent rifted continental margins (Falvey, 1974), with the lower units of the overlying Baraga Group representing deposition on the rapidly subsiding rifted margin. The upper graywacke-slate units of the Baraga Group, showing evidence of being derived from Early Proterozoic crust (Barovich and others, 1989), were probably shed from the Wisconsin magmatic terranes during their accretion to the continental margin and deposited in a foredeep or foreland basin (Hoffman, 1987).

The Wisconsin magmatic terranes consist predominantly of calc-alkaline volcanic-plutonic complexes that are interpreted to have formed as complex island arcs during subduction south of the Superior craton. Two terranes are distinguished on the basis of lithology and structure (Fig. 1). A northern terrane, the Pembine-Wausau terrane, formed as an oceanic arc over a south-directed subduction zone during the interval 1880-1860 Ma (Sims and others, 1989). This arc collided with and may have partially overridden the continental margin at about 1860 Ma, producing a fold and thrust belt to the north. A southern terrane, the Marshfield terrane, consists of remnants of mafic to felsic volcanic rocks that were deposited about 1860 Ma on Archean basement and on foliated tonalite to granite bodies ranging in age from about 1890 to 1870 Ma (Sims and others, 1989). This terrane appears to have collided with the already accreted northern arc terrane about 1840 Ma, probably by north-directed subduction. Postcollisional rhyolite and granite (1835 Ma) straddle the suture between the two arc terranes.

Collision of the Wisconsin magmatic terranes with the Superior craton margin caused the Penokean orogeny. Collision was probably oblique, with initial suturing along promontories of the continental margin. Such diachronous suturing would result in complex patterns of stress trajectories in the continental margin and accreted arcs that may account for the multiple deformation events and structural variations of the Penokean Orogen.

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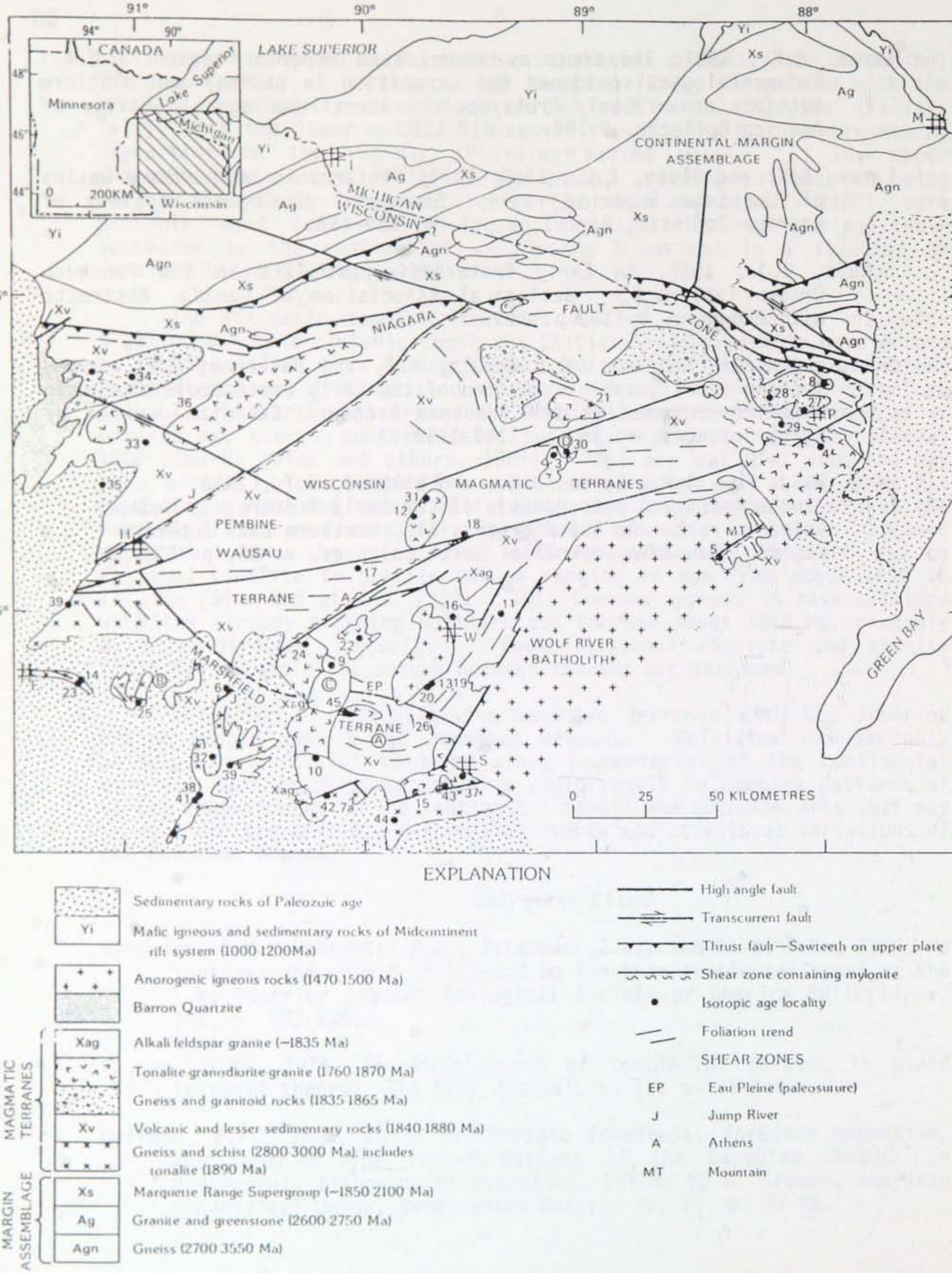


FIG. 1. Geologic map of eastern part of Lake Superior region showing relationships of rocks within Penokean Orogen. E, Eau Claire; H, Holcombe; I, Ironwood; M, Marquette; P, Pembine; S, Stevens Point; W, Wausau. Yi, Precambrian Y = Middle Proterozoic (900–1600 Ma); Xag etc., Precambrian X = Early Proterozoic (1600–2500 Ma); Ag etc., Archean (2500–3800 Ma). Compiled by P. K. Sims.

# BI-POLAR TECTONIC TRANSPORT IN THE PENOKEAN OROGEN OF NORTHERN MICHIGAN AND WISCONSIN

by

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## TECTONIC SCENARIO

It is commonly accepted that the Penokean orogen formed as the result of plate tectonic activity. Most authors agree that evolution of the orogen in Ontario, Michigan, Wisconsin, and Minnesota (fig. 1) began with a rifting phase, during which basins and troughs formed along the passive southern margin of the Archean Superior craton (figs. 1 and 2A). This was accompanied by deposition of Early Proterozoic rocks of the Marquette Range Supergroup in Michigan and equivalent Animikie Group rocks in Minnesota. Continued spreading resulted in formation of an ocean basin between the northern southern segments of the Archean craton (fig. 2B). The rifting and spreading phases were followed by a convergent phase and southward subduction of oceanic crust to form island arc-related volcanic and plutonic rocks of the present-day Wisconsin magmatic terranes (fig. 2C). These magmatic rocks were subsequently accreted (fig. 2D) to the continental margin on the north (presently northern Michigan) about 1,860 Ma (Sims and others, 1989; LaBerge and others, 1984; Larue and Sloss, 1980; Cambray, 1978). Such accretion of the magmatic rocks to the continental margin of the Superior craton resulted in northward thrusting of rocks onto the craton (Attoh and Klasner, 1989; Klasner and Cannon, 1989; Klasner and others, 1988a, 1988b; Holst, 1982, 1984). Hoffman (1987) and Southwick and others (in press) have recently suggested that the iron-bearing sequences of the Animikie Group rocks and Marquette Range Supergroup rocks were deposited in foreland basins that developed on the continental margin in front of the accretionary wedge. Neodymium (Nd) isotope studies on the Biwabik and Negaunee Iron Formations (of the Animikie Group and Marquette Range Supergroup respectively) suggest deposition of these rocks about 2,100 $\pm$ 52 Ma (Gerlack and others, 1988).

Continued convergence (fig. 2E) resulted in a similar sequence of events in what is now central and southern Wisconsin (LaBerge and others, in press; LaBerge and Klasner, 1989, 1988, 1986; and LaBerge, 1986). Northward subduction of oceanic crust, as shown in figure 2E, resulted in formation of additional volcanic and plutonic rocks of, what are now, the Wisconsin magmatic terranes. Convergence culminated in accretion of the magmatic rocks to the Archean continental crust in the south (fig. 2F). It is likely that the southern continental crust is exotic, that is, not the original part of the Superior craton that broke away during the intitial rifting phase (figs. 2A and 2B).

Accretion of the Wisconsin magmatic rocks to the southern continental margin resulted in southward thrusting of Early Proterozoic sedimentary rocks onto the southern margin.

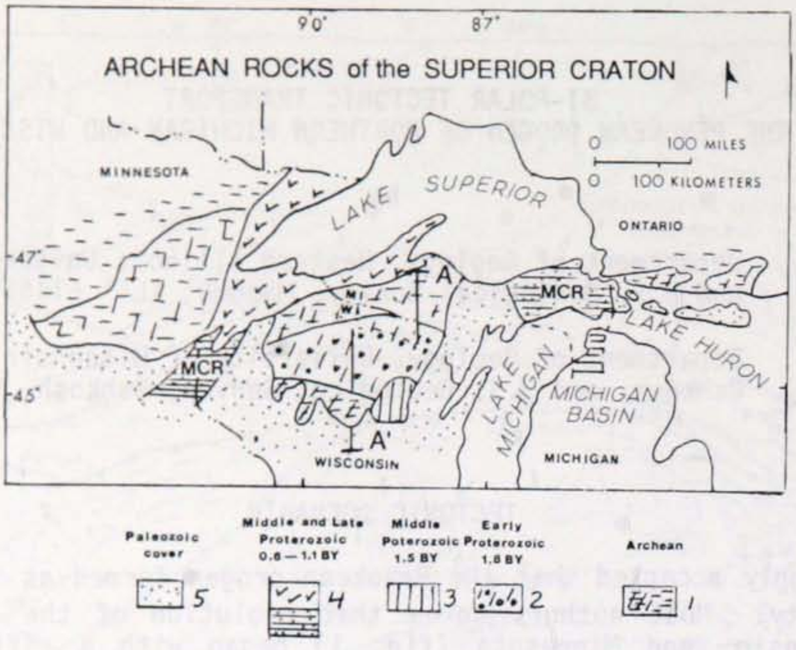



Figure 1 -- Tectonic setting of the soutern Lake Superior Region. Location of exposed rocks in the Penokean orogen is indicated by slanted hachure marks: 1, Early Proterozoic supracrustal rocks underlain by Archean gneiss and granite greenstone in Ontario, northern Michigan, Wisconsin and Minnesota; 2, magmatic terranes in northern and central Wisconsin; 3, Middle Proterozoic (1.5 Ga) anorogenic plutons; 4, Middle and Late Proterozoic rocks of the Midcontinent rift system (MCR) -- horizontal lines indicate southward extension of the MCR beneath Paleozoic cover; 5, Paleozoic cover. MI = Michigan, WI = Wisconsin, A-A[1] is location of composite profile shown in figures 2 and 7. After Attoh and Klasner (1989).

## EVIDENCE

The model for tectonic evolution of the Penokean orogen, set forth above, comes from the work of many geologists. In the following paragraphs we outline some of the scientific data that supports the model proposed above.

In southern Wisconsin (fig. 3) widely scattered areas of quartzite and locally (at Baraboo) slate, dolomite and iron-formation are interpreted to be remnants of a once extensive sedimentary sequence deposited on a passive continental margin. Archean rocks of unknown extent, locally exposed in central Wisconsin, represent the basement on which the sediments were deposited. Especially at Baraboo and at Hamilton Mounds (figs. 3, 4A, and 4B), and elsewhere in the region, structures in the quartzites indicate south-directed folding and thrusting (LaBerge and other, in press; Cambray, 1987; LaBerge and Klasner, 1986) onto the Archean rocks. We suggest that the 1,750 meter-thick sedimentary succession, along with the widespread evidence for south-verging folding and thrusting in this region, argues for collision of a substantial mass of Archean rocks -- perhaps of micro continent size -- with the Proterozoic island arc complex, which is now the Wisconsin magmatic terranes. Sims and others (1989) conclude that this collision took place about 1,840 Ma.

In the magmatic terrane of central Wisconsin (fig. 3) structures are upright-subertical to vertical (LaBerge and Myers, 1984; LaBerge and others, 1984). They have no consistent sense of structural vergence.

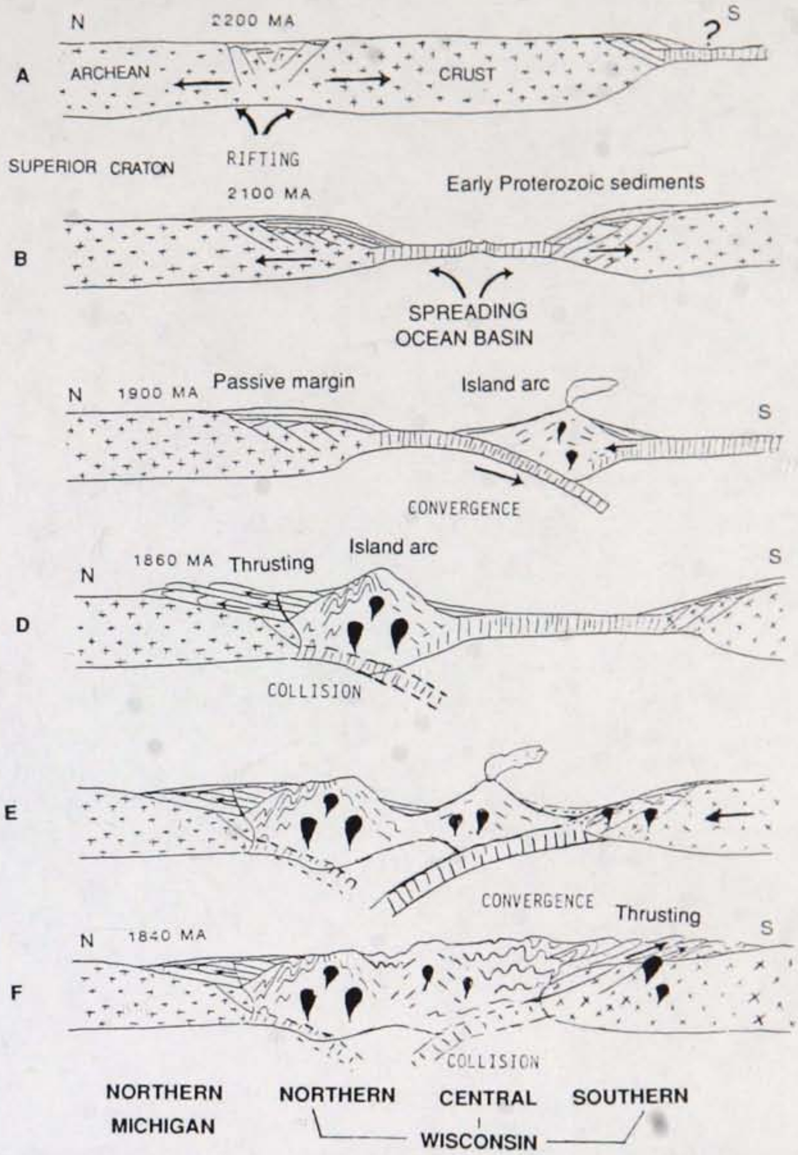



Figure 2 -- Hypothetical diagram showing the evolution of the Penokean orogen during Early Proterozoic time. This cross-sectional diagram lies along composite profile A-A[1] in figures 1 and 3.

In the magmatic terrane of northeast Wisconsin in the area of the Dunbar dome (fig. 3) structures have a strong sense of northerly structural vergence (Sims and others, 1985). Klasner and Osterfeld (1984) suggest that the 1,850 Ma Dunbar dome was tectonically transported northward and Attoh and Klasner (1989) have suggested that the magmatic rocks in the Dunbar area are para-allochthonous, having been thurst northward onto the buried continental margin (fig. 5).

Structures on the continental foreland of northern Michigan also provide strong evidence for northward structural vergence. Drag folds such as that in fig. 6A, northward overturned bedding, and north-verging thrust systems (fig. 6B) indicate northward structural transport in this region. A hypothetical cross-section across the continental margin of northern Michigan (fig. 6C) illustrates the possible nature of deformation.

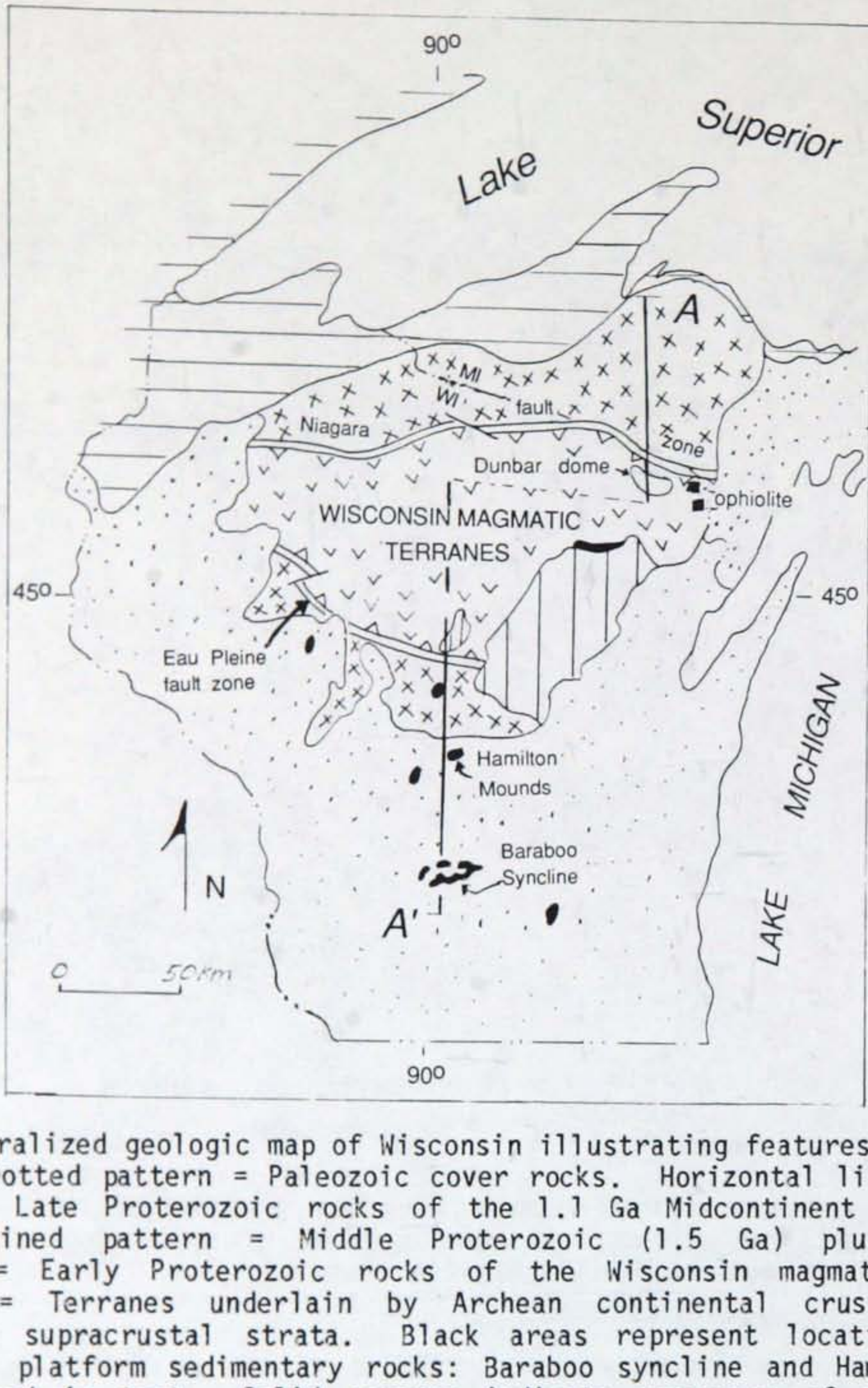
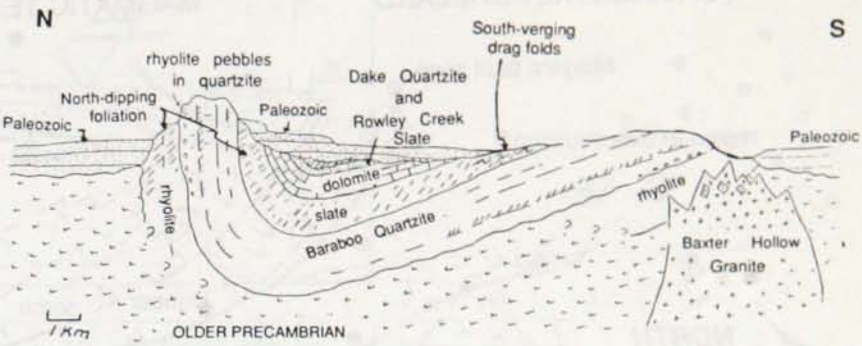



Figure 3 -- Generalized geologic map of Wisconsin illustrating features mentioned in text. Dotted pattern = Paleozoic cover rocks. Horizontal lined pattern = Middle and Late Proterozoic rocks of the 1.1 Ga Midcontinent rift system. Vertical lined pattern = Middle Proterozoic (1.5 Ga) plutonic rocks. V-pattern = Early Proterozoic rocks of the Wisconsin magmatic terranes. X-pattern = Terranes underlain by Archean continental crust and Early Proterozoic supracrustal strata. Black areas represent location of Early Proterozoic platform sedimentary rocks: Baraboo syncline and Hamilton Mounds are discussed in text. Solid squares indicate exposures of oceanic crust (ophiolite) sequences. The Niagara fault zone and Eau Pleine fault zone are the northern and southern suture zones, respectively, along which rocks of the Wisconsin magmatic terranes were accreted to the Archean continental margin rocks. Profile A-A[1] is the same as that shown on figure 1, and is the location of the diagrammiatic composite cross-sections shown in figures 2 and 7. After LaBerge and others (1989).

A




B

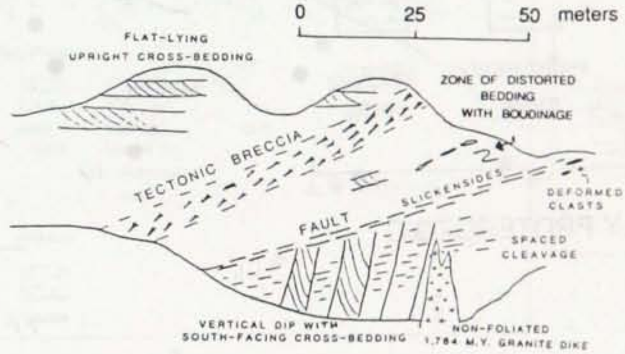



Figure 4

A) Cross-section of the Baraboo syncline (see figure 3 for location). The asymmetric fold with near vertical north limb and gently north-dipping south limb and the north-dipping axial planar cleavage indicate southward tectonic transport (southward vergence). After LaBerge and others (in press).

B) Northeast-oriented composite cross-section of quarry faces at Hamilton Mounds (see figure 3 for location). Foliation in rocks below the slickensided fault surface dips north. Bedding is overturned toward the south and drag folds above the fault zone are south-verging. A zone of tectonic breccia with tabular quartzite clasts cemented by white quartz exists above the slickensided fault surface. The clasts tend to dip north and are imbricately overlapped toward the south. All data indicate southward-vergence. After LaBerge and others (in press).

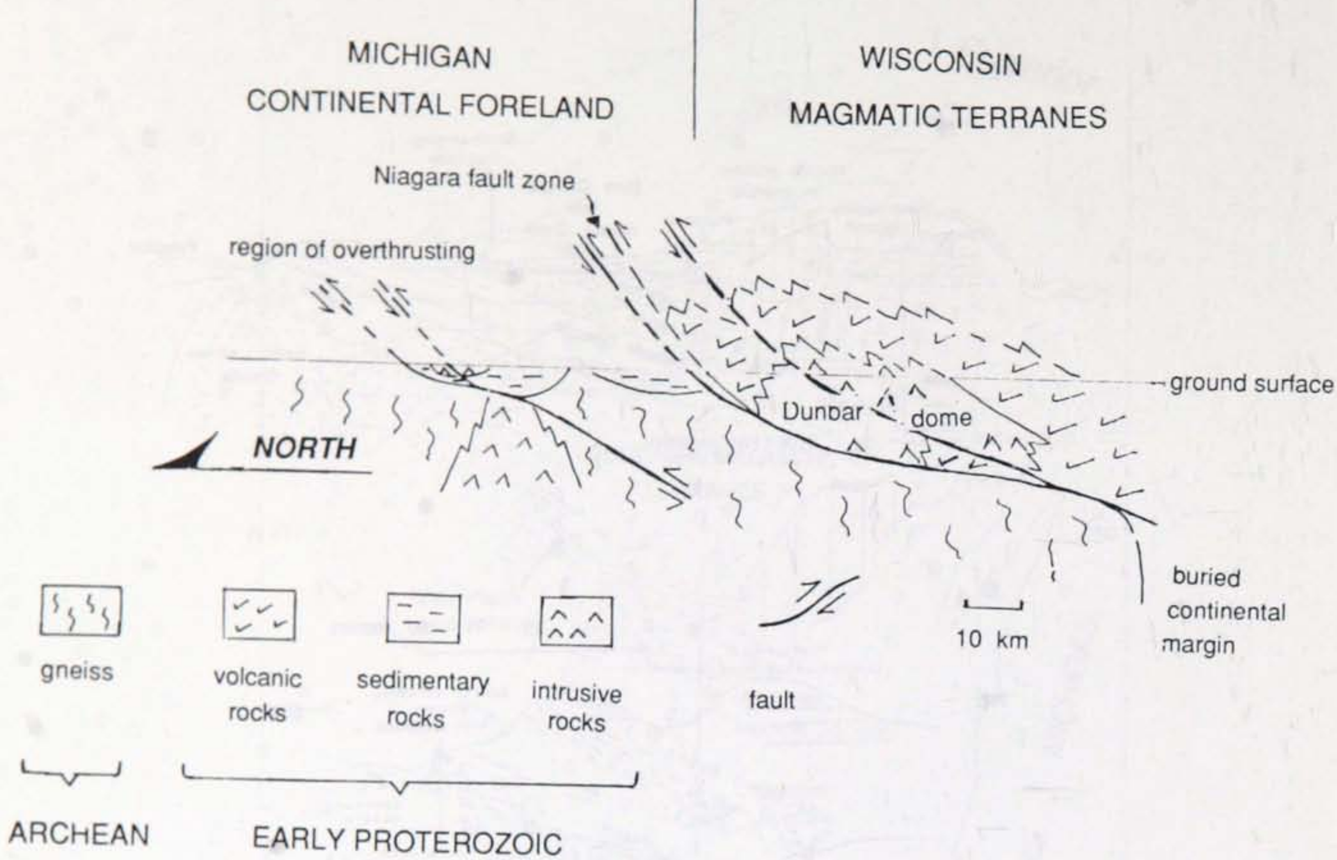



Figure 5 -- Diagram based on structural analyses and gravity modeling showing para-autochthonous nature of the Dunbar dome and Early Proterozoic overthrusting onto the Archean continental foreland of northern Michigan. The Niagara fault zone is the suture along which rocks of the Wisconsin magmatic terranes were accreted to the Archean crust. After Attoh and Klasner (1989).

## SUMMARY

Taken together, from southern Wisconsin to northern Michigan (fig. 7), the Penokean orogen currently consists of: a deformed remnant of Archean basement overlain by Early Proterozoic platform sediments deformed by south-verging structures in southern Wisconsin; a region of accreted island arc volcanic rocks with upright structures in central Wisconsin; a region of island arc magmatic and plutonic rocks with north-verging structures in northern Wisconsin; and an Early Proterozoic continental margin with platform sediments and prominent north-verging structures in northern Michigan.

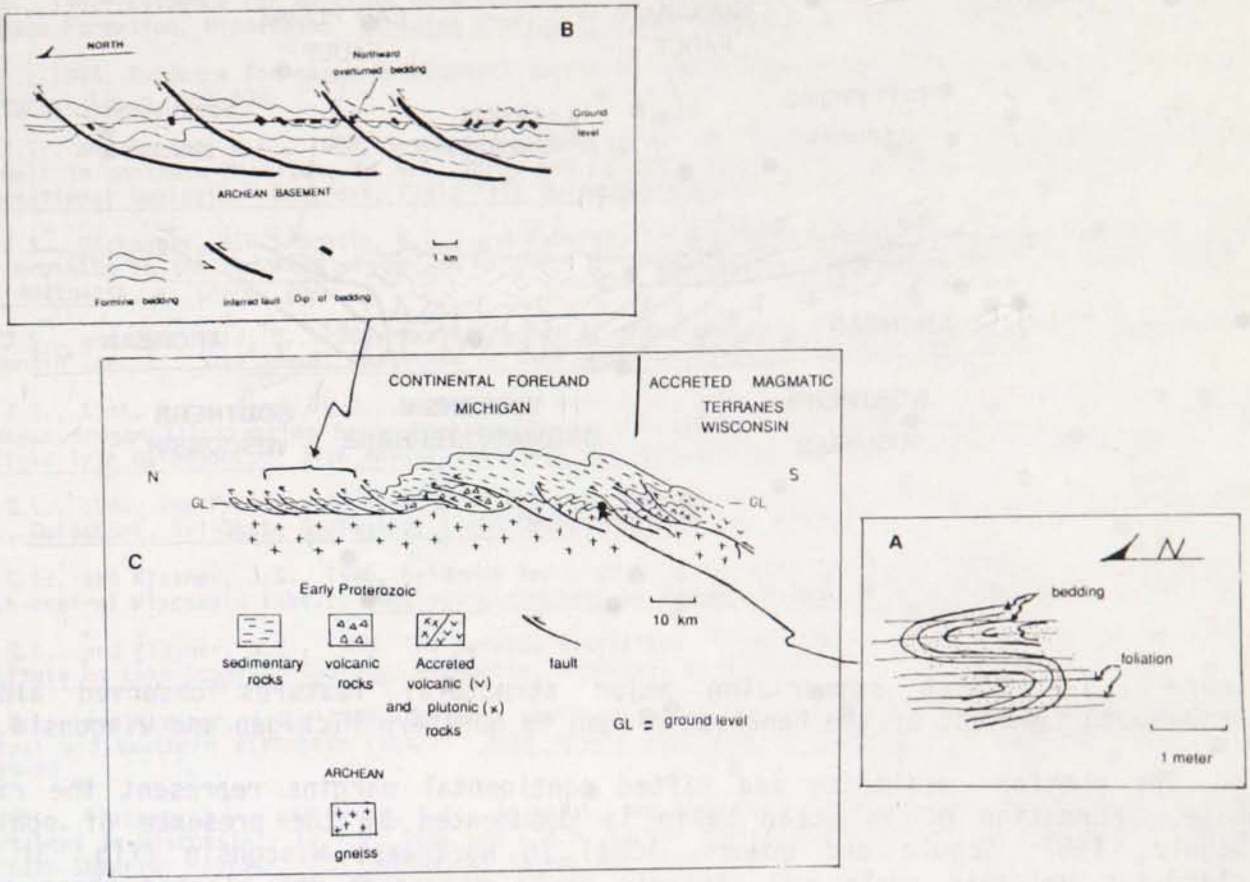



Figure 6

A) North-verging drag fold in basal quartzite of Marquette Range Supergroup rocks in northern Michigan. As shown in the diagram, foliation is nearly flat lying at this location.

B) Series of deformed strata in northern Michigan. All structures along this 16 km-long profile are north verging; beds are overturned to north in places; and foliation dips south. Faults are inferred from changes in structural orientation along profile. After Klasner and others (1988b).

C) Hypothetical cross section showing northward-verging folding and faulting on continental foreland of northern Michigan. Locations of figures 6A, 6B are shown on this section. After Klasner and others (1988b).

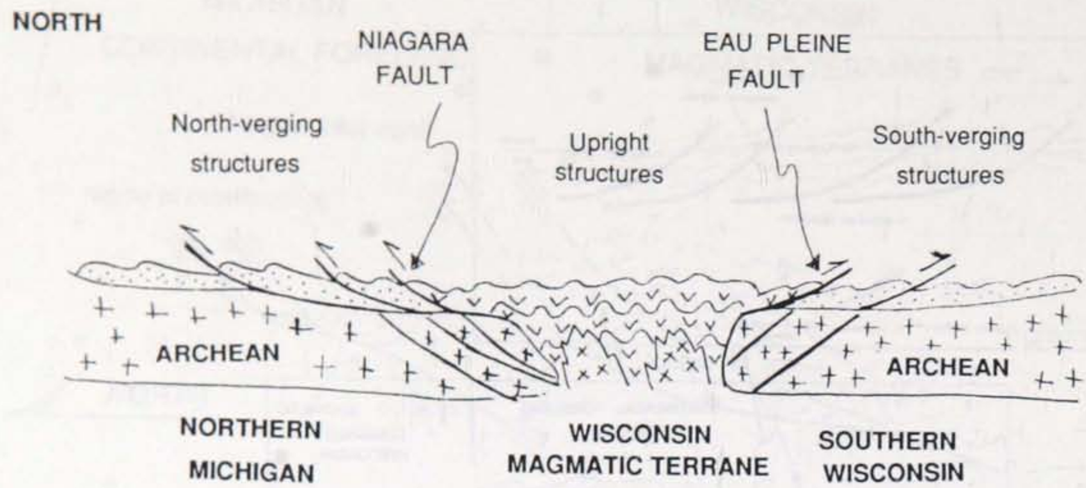



Figure 7 -- Sketch summarizing major structural features observed along a north-south transect of the Penokean orogen in northern Michigan and Wisconsin.

The platform sediments and rifted continental margins represent the rifting phase. Formation of an ocean basin is documented by the presence of ophiolite (Schulz, 1987; Schulz and others, 1984) in northeast Wisconsin (fig. 3). The island-arc volcanic rocks and plutonic rocks represent the closing phase of the ocean. Final closure resulted in accretion of magmatic rocks to the continental margins along the Eau Pleine and Niagara sutures (figs. 3 and 7) in the south and north respectively. Prominent south-verging structures formed on the continental margin in southern Wisconsin and north-verging structures on the continental margin in northern Michigan. We suggest that these structural data indicate that the Penokean orogeny was a bi-polar event.

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# STRATIGRAPHIC AND TECTONIC FRAMEWORK OF THE EARLY PROTEROZOIC PENOKEAN OROGEN IN EAST-CENTRAL MINNESOTA

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Early Proterozoic supracrustal sequences in the Lake Huron and Lake Superior regions constitute a discontinuous linear fold belt some 1300 km long, which extends from eastern Ontario into Minnesota along the southern margin of the Superior Province of the Canadian Shield (Fig. 1). These sequences compose the major part of the Southern Province and are referred to here as the Penokean fold belt or the Penokean orogen. The Early Proterozic sequences are transected at both ends of Lake Superior by the Middle Proterozoic Midcontinent rift system and, in northeastern Ontario, by the Late Proterozoic Grenville Front tectonic zone, which largely obliterates the primary features of the Early Proterozoic remnants within the Grenville Province. The Early Proterozoic rocks are overlain to the south by flat-lying Phanerozic strata.

From west to east, the supracrustal sequences consist of the Animikie, Mille Lacs, and North range groups in Minnesota; the Marquette Range Supergroup in northern Michigan and Wisconsin; and the Huron Supergroup in northeastern Ontario. These strata were deposited between 2500 and 1850 m.y. ago and were subsequently deformed, metamorphosed, and intruded by plutonic rocks. Except for the Grenville orogen, the major tectonic event affecting these Early Proterozoic successions was the Penokean orogeny (2000 - 1760 m.y.).

A possible correlation of the stratified rocks in the United States with part of the Huron Supergroup in Ontario has been debated for nearly 100 years. The basal part of the Chocolay Group at places in Michigan (Fig. 2) contains coarse clastic deposits, named the Reany Creek, Enchantment Lake, and Fern Creek Formations, which like the Huron Supergroup could be partly glacial in origin. (Morey, 1985 and references cited therein).

Isotopic ages do not preclude the suggested correlation between the lower part of the Chocolay Group and the uppermost Huron Supergroup (Morey and Van Schmus, 1988 and references cited therein). The Huron Supergroup was deposited during the interval 2480-2200 m.y.; volcanic rocks near the base have U-Pb zircon ages of approximately 2480 Ma, and the sequence is cut by Nipissing Diabase dikes that have been dated at 2219 $\pm$ 4 Ma. In the Lake Superior region, isotopic ages from the stratified rocks are poorly constrained. A U-Pb zircon age of 1910 $\pm$ 10 Ma has been reported from a sample of rhyolite in the Hemlock Formation (see Fig. 4), but no other direct ages are available. Possibly lower limits on the age of deposition are given by Sm-Nd model ages of 2120 $\pm$ 67 Ma on mafic dikes in the basement of northern Minnesota that directly underlie strata of the Animikie Group, and by an age of 1970 Ma on remetamorphosed basement rocks that underlie the upper part of the Chocolay Group in the Felch trough area, upper Michigan. These data imply that the entire Animikie Group in Minnesota and most of the Marquette Range Supergroup in Wisconsin and Michigan are distinctly younger than the Huron Supergroup in Ontario.

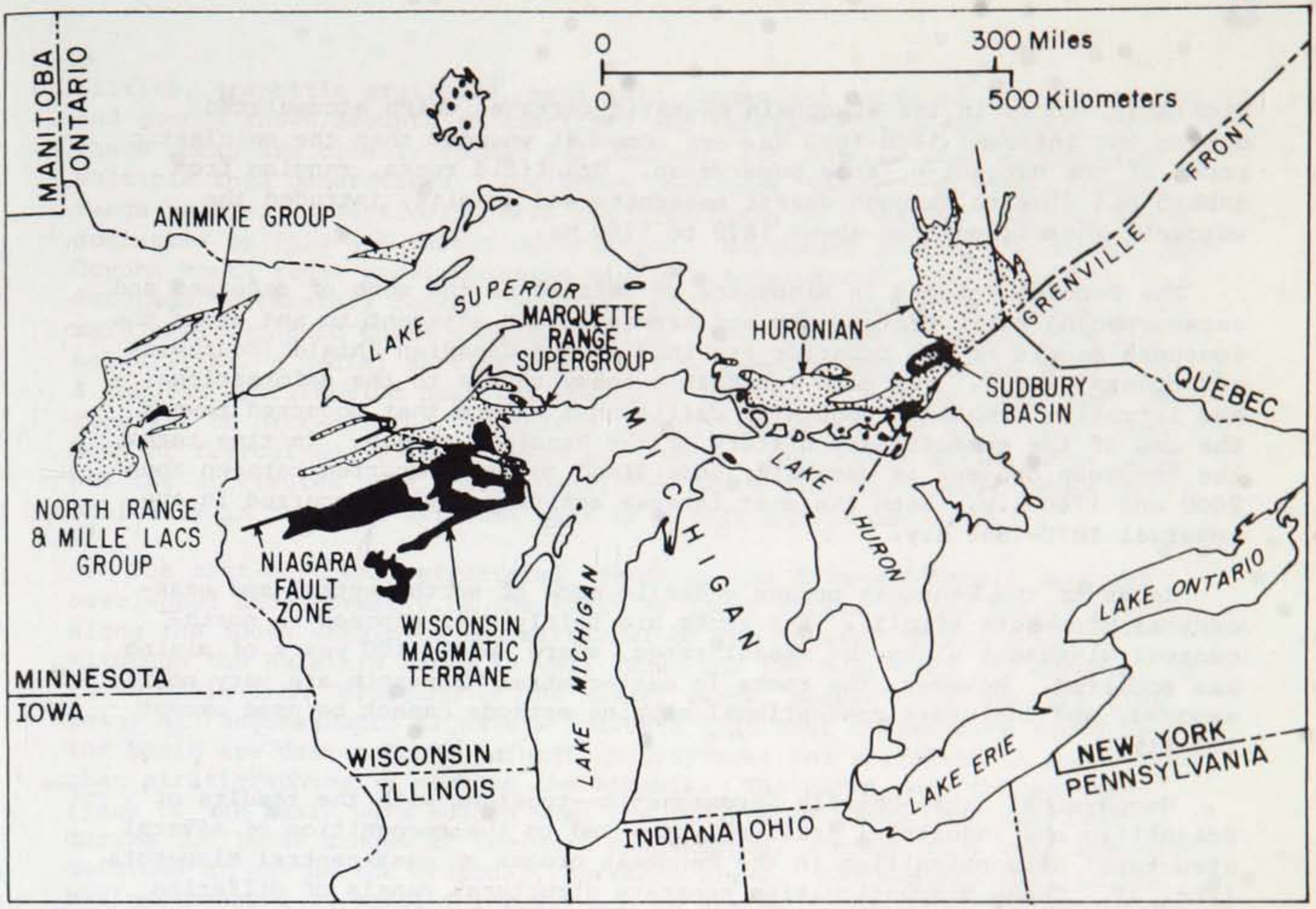


Figure 1. Simplified geologic map showing the distribution of Early Proterozoic rocks of the Great Lakes region (Morey, 1989).

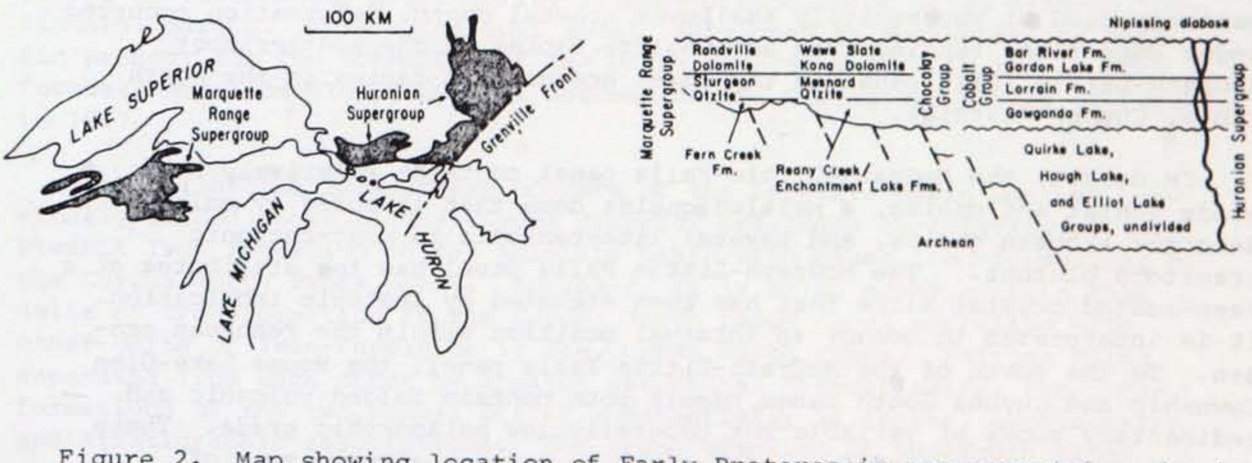


Figure 2. Map showing location of Early Proterozoic supracrustal rocks in Michigan and Ontario and a schematic representation of proposed correlations between the Huron and Marquette range Supergroups (Morey, 1989).

Similarly, rocks in the Wisconsin magmatic terrane, which accumulated during the interval 1840-1880 Ma, are somewhat younger than the epiclastic rocks of the Marquette Range Supergroup. Granitoid rocks, ranging from gabbro and diorite through quartz monzonite and granite, intruded the volcanic succession from about 1870 to 1760 Ma.

The Penokean orogen in Minnesota is defined as the zone of deformed and metamorphosed Early Proterozoic and Archean rocks adjacent to and along the southern margin of the Superior Province of the Canadian Shield (Southwick and others, 1988). The term Penokean orogeny refers to the deformational and intrusive events of generally collisional nature that occurred toward the end of the evolutionary history of the Penokean orogen. In time terms, the Penokean orogeny is generally understood to have occurred between about 2000 and 1760 m.y., with the most intense activity having occurred in the interval 1870-1850 m.y.

Rocks of the Penokean orogen underlie much of north-central and east-central Minnesota (Fig.3). The rocks are fairly well exposed in north-central Minnesota along the Mesabi range, where nearly 100 years of mining has occurred. However, the rocks in east-central Minnesota are very poorly exposed, and therefore conventional mapping methods cannot be used except locally.

Geophysical data--chiefly aeromagnetic--together with the results of scientific and industrial drilling, have led to the recognition of several structural discontinuities in the Penokean orogen of east-central Minnesota (Fig. 3). These discontinuities separate structural panels of differing stratigraphy, structure style, and metamorphic grade, and appear to be zones of thrust faulting (Southwick and others, 1988). Regionally, the thrust-faulted, tectonically imbricated terrane constitutes a fold-and-thrust belt that was emplaced onto the southern margin of the Superior craton early in the deformational history of the Penokean orogen. Imbricate thrusting and recumbent folding may have begun as early as the time interval 2130-2180 m.y. and continued episodically for some time thereafter. From south to north, the thrust-bounded structural panels contain rocks metamorphosed at successively shallower crustal depth; deformation occurred under conditions ranging from amphibolite facies in the southernmost McGrath-Little Falls panel to the lower greenschist facies in the North range, Cuyuna district.

In detail, the McGrath-Little Falls panel contains relatively high grade schist and gneiss, a mantled gneiss dome that is cored by multiply deformed Archean gneiss, and several late-tectonic to post-tectonic granitoid plutons. The McGrath-Little Falls panel has the attributes of a deep-seated crustal slice that has been elevated by tectonic imbrication. It is interpreted to occupy an internal position within the Penokean orogen. To the north of the McGrath-Little Falls panel, the Moose Lake-Glen Township and Cuyuna South range panels both contain folded volcanic and sedimentary rocks of variable but generally low metamorphic grade. These two panels are separated from each other by a long, arcuate zone of probable thrusting that is localized by a belt of structurally weak and highly deformed graphitic schist. Both panels contain considerable mafic to intermediate volcanic and hypabyssal rock, abundant metapelite, metasiltite, graphitic argillite, many thin, lensoidal units of iron-formation, and poorly known amounts of quartzite and related arenaceous rocks. All these rocks are closely folded and cleaved, and locally show evidence of multiple fold generations. The Moose Lake-Glen Township and Cuyuna South range panels, collectively, have the attributes of a medial tectonic zone dominated by fold-and-thrust deformation. Northwest of these terranes, the Cuyuna North range panel contains weakly metamorphosed, less strongly deformed sedimentary rocks. Volcanic rocks are volumetrically minor. The main stratigraphic units include a thick lower section of metapelite and metasiltite, a medial section dominated by iron-formation units that define a complex synclinorium near the center of the panel, and an upper unit that consists of dark-colored, graphitic argillite and siltite with local interbeds of ferruginous chert. Taken as a whole, the Cuyuna North range panel has the attributes of a small restricted basin that was incorporated tectonically in the more external part of a fold-and-thrust belt.

The east-northeast structural trends of the fold-and-thrust belt are overlapped unconformably on the north by lower strata of the Animikie Group along the south margin of the main bowl of the Animikie basin (Fig. 3). Although the Animikie basin is well known for the huge iron ore deposits of the Biwabik Iron Formation of the Mesabi range, sparse outcrops, drilling data, and aeromagnetic signature indicate that most sedimentary rocks in the basin are dark-colored turbiditic graywacke and argillite in formations that stratigraphically overlie the Biwabik. The rocks near the southern flank of the basin were folded when the underlying rocks were refolded during the later stages of regional compression, whereas they were scarcely deformed at all on the cratonal, northern flank. Sedimentary fill in the basin decreases in total thickness and in degree of low-grade metamorphism from southeast to northwest. Primary sedimentary structures in rocks near the rim on the north side of the basin clearly indicate a northern (cratonal) source, whereas the sedimentological and geochemical attributes of at least some of the lithic graywacke near the southern margin of the basin are consistent with a southern provenance. Taken as a whole, the broad features of the Animikie basin and its smaller analog are consistent with those of a migrating foredeep produced by tectonic loading and downbowing of continental crust during attempted subduction of an Archean continental margin. This conceptual model, developed by Southwick and Morey (in press), for the formation of the basin accords well with the general foredeep model developed by Paul Hoffman of the Geological Survey of Canada in 1987.

The unconformable southern contact of the Animikie basin against previously folded rocks of the Cuyuna district precludes correlation of the Biwabik Iron Formation in the Animikie basin with other iron-formations of the Cuyuna North range panel. Moreover, it is by no means certain that units of iron-formation in the Cuyuna North range panel correlate with other units of iron-formation in the Cuyuna South range. Besides being separated from each other by a major structural discontinuity, the iron-formations of the two panels differ substantially in facies, geochemistry, and stratigraphic associations. The iron-formation of the South range panel is mainly of sulfide, carbonate, and silicate facies and was deposited as lenticular masses and thin layers in close stratigraphic proximity to mafic volcanic rocks and euxinic black shale. It is akin to Algoma-type iron-formation as defined by Gordon Gross of the Geological Survey of Canada. In contrast, the iron-formations of the Cuyuna North range panel have a more blanket-like morphology, are interbedded with dark-colored argillite and siltite, and are predominantly of carbonate and oxide facies. Both they and the iron-formation units of the Animikie basin have sedimentological attributes typical of Lake Superior-type iron-formation as defined by Gordon Gross.

The inference that there are three stratigraphically important iron-formation in Minnesota has implications for correlations with Early Proterozoic rocks in Wisconsin and Michigan (Fig.4). The stratified rocks of the Marquette Range Supergroup that crop out on the south side of Lake Superior have attributes similar to those in Minnesota, and the two sequences have generally been correlated, mainly using a single unit of iron-formation. This work is summarized in the now classical studies of H.L. James of the U.S. Geological Survey which were published in the late 1950s and 1960s. We now believe that there are broad similarities between the Menominee Group of Michigan and the North Range group in east-central Minnesota. Both are terranes of tectonically imbricated "continental" rocks that abut large turbidite basins on the north where they unconformably overlie Archean cratonic basement. Furthermore, the geologic attributes of the Animikie and Baraga Groups are compatible in both cases with a foredeep depositional setting. The age of the Baraga-Animikie depositional assemblage has not been established. However, isotopic studies of mafic volcanic rocks from the Glen Township Formation in Minnesota yield an age of 2197 ± 39 Ma, or some 300 m.y. older than the 1910 ± 10 Ma age from geochemically similar basalts of the Hemlock Formation of upper Michigan. Thus the interval from 2197 ± 39 Ma to 1910 ± 10 Ma may represent the span of deposition for Early Proterozoic strata in the Lake Superior region that are older than the Animikie-Baraga succession of Figure 4. In Minnesota, deformation, metamorphism, and plutonism culminated about 1870 m.y. when voluminous syntectonic plutons were emplaced (Morey and Southwick, in press). Crustal downflexure and foredeep development may have begun prior to the onset of extensive plutonism and persevered for a long period of time. Early foredeeps--such as the ancestor to the Cuyuna North range synclinorium--eventually were overridden and incorporated in the imbricate thrust stack. Later foredeeps, including the Animikie basin, followed in turn as subduction continued and then waned. A second regional episode of deformation and metamorphism accompanied the emplacement of late-tectonic to post-tectonic intrusions in the interval 1820(?) to 1770 Ma; structures primarily of this generation developed in rocks of the Animikie Group and are overprinted on earlier structures in older rocks. Isostatic uplift is inferred to have culminated by about 1740 Ma.

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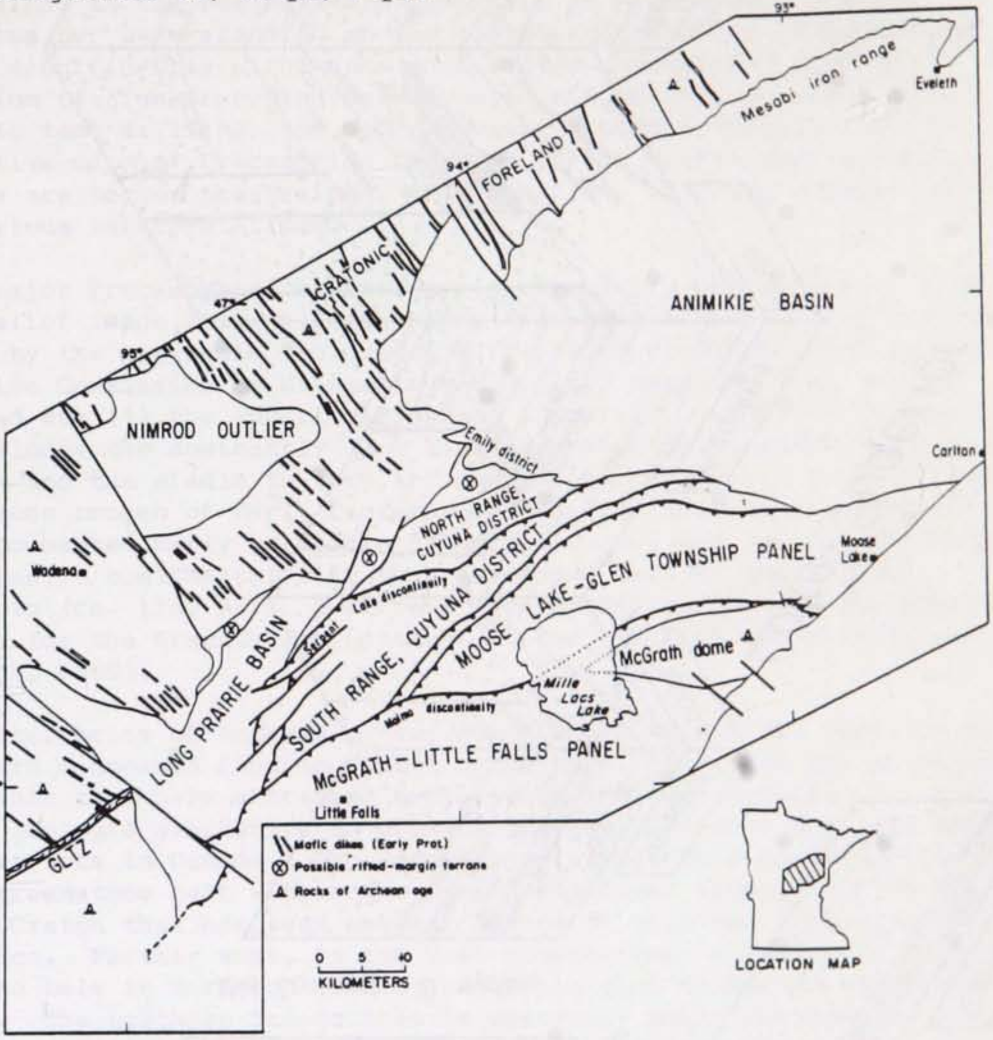


Figure 3. Generalized tectonic map of the Penokean orogen of east-central Minnesota. A fold-and-thrust belt of Early Proterozoic age consists of the McGrath-Little Falls, the Moose Lake - Glen Township, and the Cuyuna South range structural panels; the panels are bounded by structural discontinuities inferred to involve significant thrusting. The North range of the Cuyuna district is interpreted as an early foredeep that has been tectonically incorporated in the external zone of the fold-and-thrust mass. A major foredeep of Early Proterozoic age, but consequent to some and perhaps much deformation in the fold-and-thrust belt, consists of the Animikie basin, the Long Prairie basin, and the Nimrod outlier. The sedimentary fill of these basins rests unconformably on cratonic basement of Archean age along the northwest margins of the basins (Southwick and others, 1988).

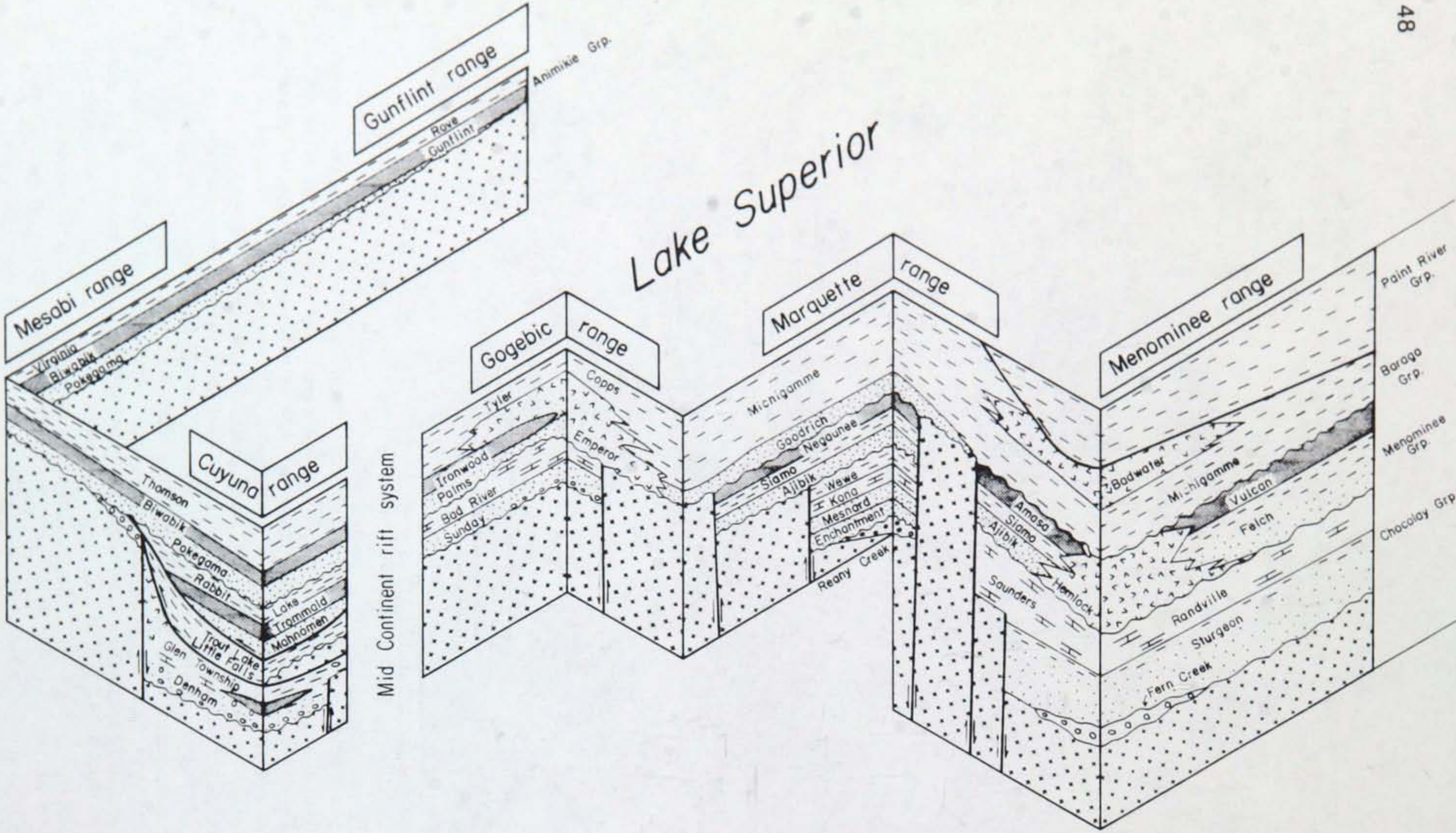


Figure 4. Fence diagram showing regional correlations of Early Proterozoic strata in the Lake Superior region.

PRECAMBRIAN GEOLOGIC FRAMEWORK IN MINNESOTA

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Much of Minnesota is underlain by bedrock of Precambrian age. However most of that is covered by a thick mantle of surficial materials of Pleistocene and Holocene age. Consequently conventional mapping at a scale of 1:24,000 or 1:48,000 can be accomplished in only about 15 percent of the state, mainly in the northeastern quadrant, or "Arrowhead" region. Nonetheless our understanding of the state's Precambrian framework has improved significantly within the past decade through the combined application of high-precision aeromagnetic and gravity surveys, follow-up scientific test drilling, and outcrop-based geologic mapping. The revised interpretive maps of Precambrian terranes in the drift-covered parts of Minnesota are better constrained, more detailed, and more sophisticated than previous interpretations.

The major Precambrian terranes are particularly well shown in a new shaded relief image (Fig. 1) made using high-resolution aeromagnetic data acquired by the Minnesota Geological Survey with funds provided by the Legislative Commission on Minnesota Resources. Terranes that can be recognized are (1) the Superior Province (Superior Craton) of Archean age, which includes the dominantly late Archean greenstone-granite terrane in the north and the middle to late Archean gneiss terrane in the south; (2) the Penokean orogen of Early Proterozoic age; (3) the Sioux Quartzite of poorly documented Early or Middle Proterozoic age; and (4) the Midcontinent rift system, a continental rift that developed late in the Middle Proterozoic (ca. 1100 Ma). A current interpretation of this lithotectonic framework for the Precambrian basement of the state is shown in Figure 2 (Southwick, 1989).

To date, belts of Archean metavolcanic rocks within the Superior Craton in northern Minnesota (the so-called greenstone belts) and the major fault zones within them have attracted exploration interest because of their geologic similarity to producing greenstone-belt gold and base-metal districts in Canada. The Vermilion district is a moderately well exposed greenstone belt within the Wawa-Shebandowan subprovince of the Superior Craton that has seen several cycles of gold and base-metals exploration. Farther west, in the same subprovince, a less well exposed greenstone belt in northern Itasca County has also attracted exploration interest. The northern Itasca area is currently being remapped by Jirsa and others (1990) of the Minnesota Geological Survey under a combination of geophysical, drilling, and mapping initiatives, and much geological detail is emerging. Among other things, they have established the importance of regional-scale sinistral structures that were subsequently fragmented by mostly right-lateral fault zones oriented oblique to the strike. "Destraining" the major fault displacemnts has led to the recognition of a major break that joins the rocks of Itasca County and Vermilion district and divides them into a northern and a southern segment. Although the petrogenetic significance of this break remains to be determined, it is clearly an early feature of considerable stratigraphic importance (Fig. 3).


Figure 1. Shaded relief aeromagnetic image of Minnesota exclusive of the southeastern quarter underlain by Paleozoic rocks. Synthetic illumination is from the northwest at an inclination of 45°. Data were acquired over the past 10 years by the Minnesota Geological Survey with funds provided by the Legislative Commission on Minnesota Resources. The image was produced using the computer facilities of the Minnesota Land Management Information Center.

The rocks of the Wawa-Shebandowan province and the Wabigoon subprovince just to the north are juxtoposed along the Vermilion fault zone, a major right-lateral structure that formed late in the history of the area. The southern part of the Wabigoon subprovince comprises a strongly deformed greenstone belt that crosses the far northwestern corner of the state. It has been explored several times for both base metals and gold, and is currently being investigated under the Conterminous United States Mineral Appraisal Program (CUSMAP) of the U.S. Geological Survey, a program which studies the mineral potential of 1° x 2° quadrangles. Other geophysically identified greenstone belts in the westward extensions of the Wawa-Shebandowan and Wabigoon subprovinces lie beneath prohibitively thick overburden and have attracted little exploration interest. We know very little about these areas, and they are shown as Superior Craton, undivided on Figure 2.

An intriguing and little-understood subunit of the greenstone-granite terrane is the so-called "quiet zone" (Fig. 2). Geophysical expressions over this area are relatively flat and featureless, comparable to those found over metasedimentary belts elsewhere in the Superior Craton, and yet the drill reveals a varied geology that includes volcanic and plutonic as well as sedimentary protoliths. Many of the drill samples from the quiet zone show evidence of late-stage epidote-chlorite-albite alteration, and it may be that the featureless aeromagnetic expression is due in part to a regional episode of retrograde metamorphism in which magnetite was consumed.

The Archean gneiss terrane of southwestern Minnesota consists predominantly of quartzofeldspathic gneisses and younger granitoid intrusions that have undergone a long and eventful Precambrian history. Although relatively minor, there are gneissic protoliths of volcanic, pelitic, and iron-formation compositions that may be analogous to greenstone-belt assemblages. Current studies have established that the gneiss terrane consists of at least three distinct strata-tectonic blocks that are bounded by zones of faulting and ductile shear (Schaap, 1989). Geophysical modeling indicates that the bounding shear zones, as well as the internal structures of the Benson, Montevideo, and Morton blocks, consistently dip at low to moderate angles to the north. These regional shear zones are parallel to the Great Lakes tectonic zone, which is a probable paleosuture between the gneiss terrane on the south and the greenstone-granite terrane on the north. All of these structures reflect a major shear event in late Archean time. The Great Lakes tectonic zone also was active in Early Proterozoic time.

Linear magnetic anomalies with northwest trends that transect the Archean greenstone-granite and gneiss terranes (Fig. 1) delineate dikes of the 2125-Ma Kenora-Kabetogama swarm, the largest known dike swarm in the United States. The dikes are truncated at the structural front of the 2200 - 1760 m.y. Penokean orogen.

The Early Proterozoic Penokean orogen of east-central Minnesota is interpreted to consist of an allochthonous fold-and-thrust belt on the southeast and one or more tectonic foredeeps on the northwest (Southwick and others, 1988). The fold-and-thrust mass includes an internal zone that

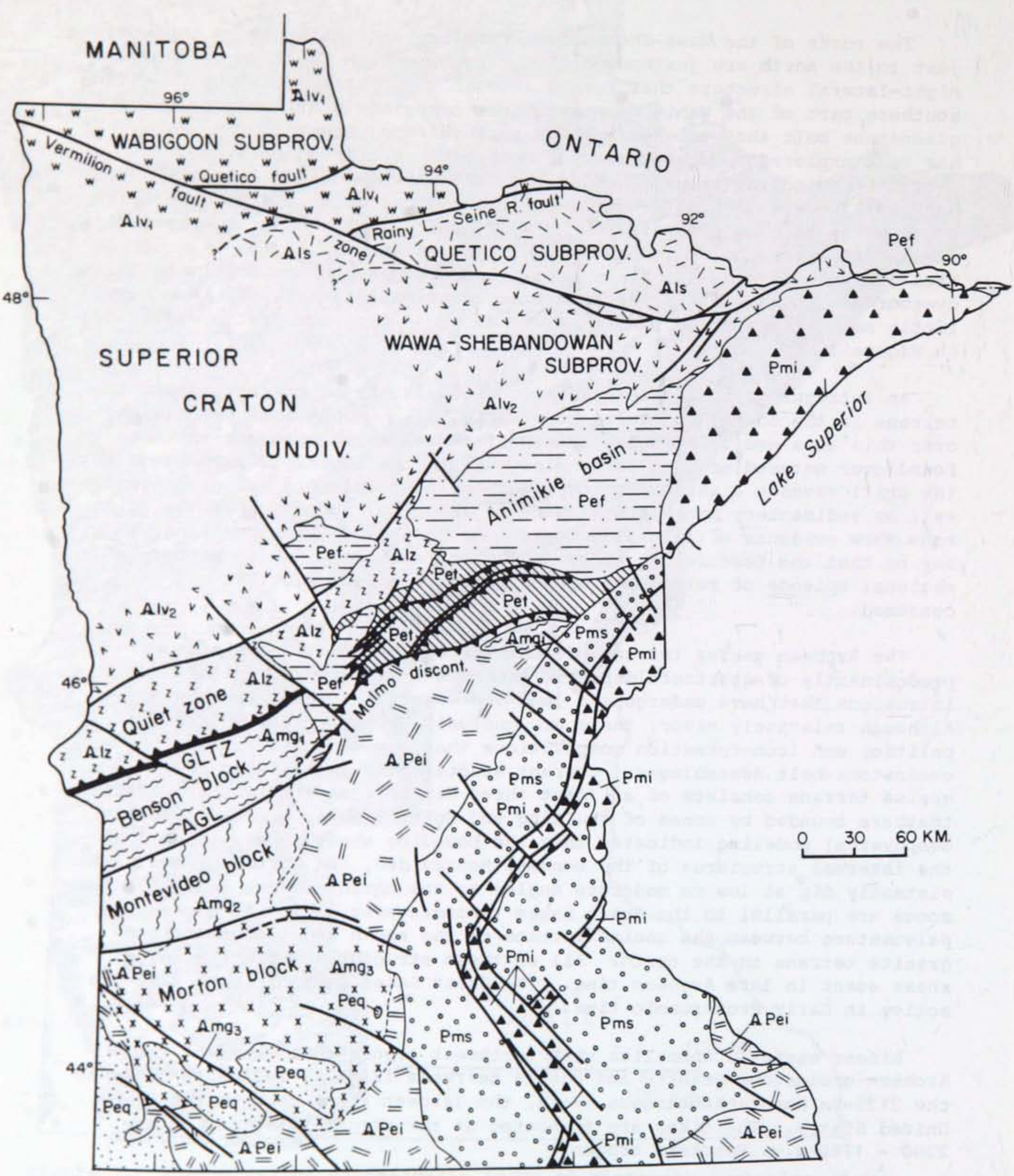


Figure 2. Simplified tectonic map of Minnesota compiled from published sources and unpublished work in progress at the Minnesota Geological Survey. Age data tabulated in explanation are from published sources except as noted. Ages from named Archean subprovince are from Canadian sample suites (Southwick, 1989).

# EXPLANATION

## MAJOR PRECAMBRIAN TERRANES OF MINNESOTA

| TECTONIC ELEMENT | | PRINCIPAL ROCK TYPES | AGE |
|---|---|---|---|
| **Midcontinent rift system** | | | |
| late- and post-rift | Pms | Fluvial and lacustrine clastic sedimentary rocks | Middle Proterozoic; 1100--1050 Ma |
| syn-rift | Pmi | Basalt, rhyolite, gabbroic intrusions; minor interflow sedimentary deposits | |
| **Sioux Quartzite basins** | Peq | Fluvial, sand-dominated redbed sequences in basins that may be fault-controlled | Early Proterozoic; probably between 1760 and 1630 Ma |
| **Penokean orogen** | | | |
| foredeeps | Pef | Turbiditic graywacke-shale sequences | |
| fold-and-thrust belt | Pet | Passive-margin metavolcanic and meta-sedimentary rocks, tectonically imbricated | Early Proterozoic; mainly between 2200 and 1760 Ma |
| intrusion-dominated magmatic terrane | APei | Syn- to post-kinematic intrusions of granitoid rocks into complex metamorphic terrane | |
| **Superior craton** | | | |
| Greenstone-granite terrane | | | |
| Wabigoon subprovince | Alv1 | Arc-like volcanoplutonic sequences; syn- to post-kinematic granitoid intrusions | Late Archean; volcano-plutonic belts mainly 2750-2695 Ma; Quetico belt 2690-2650 Ma |
| Quetico subprovince | Als | Turbidite-dominated metasedimentary rocks (accretionary complex?); granitoid intrusions | |
| Wawa-Shebandowan subprovince | Alv2 | Arc-like volcanoplutonic sequences; syn- to post-kinematic granitoic intrusions | |
| "quiet zone" | Alz | Poorly known belt of rocks comparable to Wawa-Shebandowan; regionally retrograded | Syntectonic plutonism, 2688 Ma (unpub. data, Z.E. Peterman) |
| Gneiss terrane | | | |
| Benson block | Amg1 | Poorly known terrane composed of gneiss and abundant granitoid intrusions | Middle and late Archean; complex history spans interval 3550 to 2550 Ma |
| Montevideo block | Amg2 | Amphibolite- to granulite-grade gneiss of plutonic and supracrustal derivation; granitoid intrusions | |
| Morton block | Amg3 | | |

inferred sequence of tectonic accretion

**Major structural discontinuities**

Malmo discontinuity (Early Proterozoic): Separates supracrustal panels of Penokean fold-and-thrust belt from deeper crustal zone to south

Vermilion fault zone (late Archean): Obliquely cuts and displaces subprovince boundaries within the Superior craton

Great Lakes tectonic zone (GLTZ; late Archean with probable Proterozoic reactivation): Separates high-grade gneissic terranes at southern margin of the Superior craton from classic greenstone-granite terrane of lower metamorphic grade on the north

Appleton geophysical lineament (AGL; late Archean with probable Proterozoic reactivation): Separates Benson and Montevideo blocks in gneiss terranes

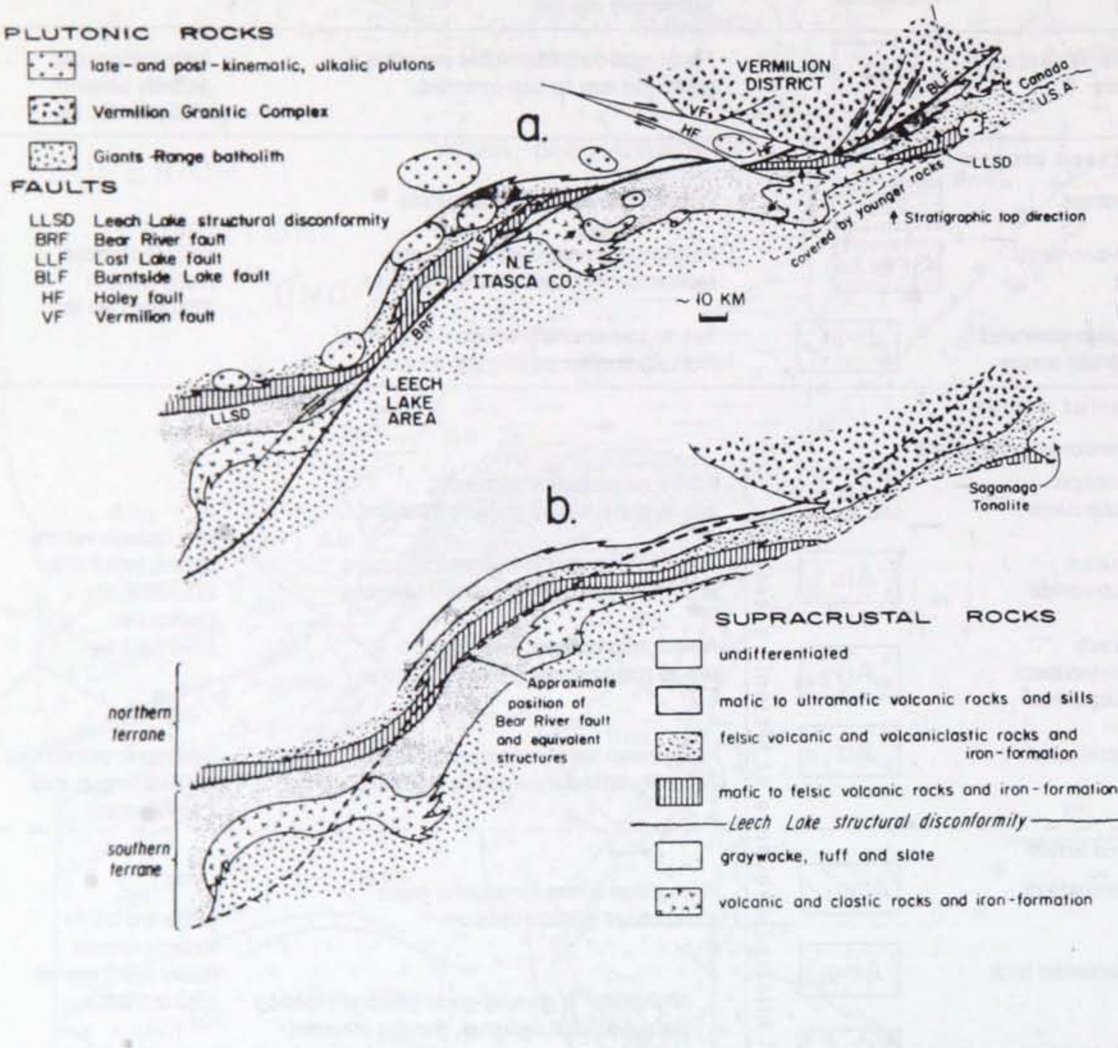


Figure 3. Simplified geologic map (a) of part of the Vermilion greenstone belt in the Vermilion district in northeastern Itasca County; and (b) reconstruction to situation before major strike-slip displacement (Jirsa and others, 1990).

is mainly gneiss and moderately high-grade pelitic schist permeated by granitoid plutons, and a medial to external zone that is mainly supracrustal rocks of moderate to low metamorphic rank. Deformed volcanic rocks constitute 20 percent or more of the medial zone of the orogen, and the possibility of finding vein-type gold and base-metal ore deposits in this environment deserves serious consideration. The latest and largest of the fore deep basins--the Animikie basin of Figure 2--contains the Biwabik Iron Formation of the Mesabi range in northern Minnesota. The range has produced over 3 billion tons of iron ore and taconite concentrates since the discovery of ore in 1890. Post-kinematic "Penokean" plutons extend southward from central Minnesota to Iowa, and together with supracrustal remnants appear to form a quasi-continuous belt around the east, south, and west sides of the Archean gneiss terrane. The tectonic and economic implications of this Proterozoic envelope about an Archean core are unresolved and unevaluated.

The Sioux Quartzite is a dominantly red-bed sequence that rests unconformably atop the Archean gneiss block and its obscure Proterozoic fringe in southwestern Minnesota. Sediment was transported by fluvial process chiefly from the northwest, off the stabilized post-Penokean craton, and was deposited in several elongate basins controlled by northwest-trending faults. Recent paleomagnetic studies (Chandler and Morey, 1990) imply that the Sioux was deposited during the interval from 1700 to 1650 m.y., which corresponds in part with the time of development of the so-called Central Plains orogen. Rocks of the Central Plains orogen are entirely covered by Phanerozoic strata of the northern Midcontinent (Fig. 4).

The rocks of the Midcontinent rift system consist dominantly of syn-rifting basaltic flows and intrusions and post-rifting clastic sediments. Prominent among the rift-related intrusive rocks is the Duluth Complex, a very large multiple intrusion that consists dominantly of troctolitic and anorthositic units. The Duluth Complex has attracted considerable attention as a potential source of copper, nickel, vanadium, titanium, cobalt, and platinum-group elements (PGEs). Non-economic occurrences of these metals have been found in several different rock types and structural settings within the complex, and exploration for PGEs is particularly lively at the present time. Rift-related sedimentary rocks include earlier sequences that are chiefly of intra-rift derivation and later sequences that were derived from cratonal souces external to the rift proper. The basins of extra-rift sedimentary rocks that flank and locally overlie the rift axis are potential habitats for alluvial gold deposits, but have never been systematically evaluated.

As noted previously, understanding the Precambrian geology of Minnesota is impeded by a thick layer of overburden. The Quaternary glacial deposits are thicker than 30 m (100 ft) almost everywhere in the state except the northeastern quadrant, and thicknesses in excess of 200 m (660 ft) are well documented above the Archean greenstone-granite terrane in the west-central and north-central regions. Much of the thickest glacial drift consists of water-saturated sand and gravel. In addition to glacial materials, the overburden in much of the state includes several tens of meters (locally more than a hundred meters) of saprolitic regolith, which developed on crystalline rock during tropical weathering in pre-Late Cretaceous time, and local erosional remnants of poorly lithified Late Cretaceous shale, siltstone, and sandstone. The total thickness of unconsolidated material above sound Precambrian basement, including saprolite, Cretaceous strata, and glacial drift, is routinely greater than 125 m (410 ft) in large areas of western Minnesota, and in places is more than twice that.

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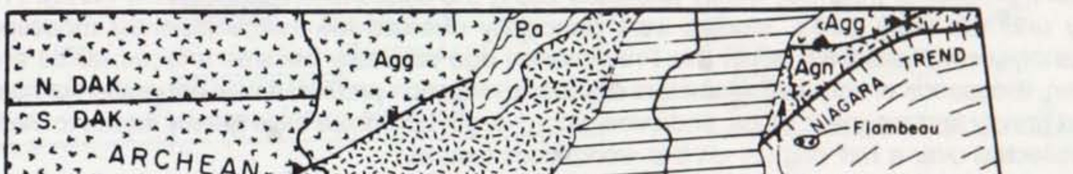


Figure 4. Generalized geologic map of the central plains region showing the relationship of the Sioux Quartzite to the Central Plains orogen (Southwick and others, 1986; Chandler and Morey, 1990).

# Structure, Stratigraphy and Economic Geology of the Proterozoic (Middle and Late Riphaen) Midcontinent Rift System, Central United States of America

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The Midcontinent Rift System (MRS) is the best documented, most studied, and best known of any rift or rift system of the United States (Dickas, 1989). The overall petrographic, stratigraphic and structural characteristics of this feature indicate it should be classed within the intracontinental (continental) category of rift zones, according to the scheme developed by Milanovsky (1978). Its subsurface and outcrop extent has been mapped over a distance of more than 2,000 km, extending from central Kansas northeast to the Lake Superior basin, and then southeast to the lower peninsula of Michigan (Figure IA). Woollard (1943) discovered the gravity high associated with this rift as a result of his transcontinental geophysical survey, but it was Thiel (1956) who established the relation between this gravity signature and causative geologic structure. After the association of this "midcontinent gravity high" with rift tectonics, several names were proposed, the most widely accepted being the Midcontinent Rift System (MRS). From its discovery until the early 1980's, studies were principally directed toward geopotential modeling and outcrop analyses. Since 1983, when this Precambrian age structure became a focus for oil and gas exploration, thousands of acquired kilometers of seismic reflection profiles, a new generation of computer enhanced gravity and magnetic maps, and three record-depth boreholes have greatly added to the wealth of data collected over a half century on this world-class feature.

Almost all of the early field work associated with this rift was conducted in the classic Keweenawan-age outcrop terrane of the Lake Superior region. Studies in this region by Paces and Davis (1988) have shown that rifting was initiated through crustal extension faulting and extrusion of volcanic rocks during the time period 1,109 to 1,094 Ma (Middle Proterozoic). These fissure-type flows accumulated, in plateau geometries, within eight overlapping basins (Green, 1982) and are locally associated with the eroded remnants of two composite shield volcanoes (Annells, 1974: Kopydlowski, 1983). After cessation of volcanism, deposition began of a thick section of clastic rocks, composed of two groups (Figure IB). The older Oronto Group, conformable with underlying volcanic strata, is composed of immature, volcanoclastic conglomerate, sandstone, and shale deposited within a transgressive-regressive regime ranging spatially and time-wise from alluvial fan to lacustrine to fluvial in depositional environment (Daniels, 1982). The younger Bayfield Group is composed of predominate mature sandstone lithology and signifies a change from an era of extension tectonics, represented by the basement volcanics and Oronto Group, to an era of compressional tectonism (Morey and Ojakangas, 1982).

In the type-section Keweenawan outcrop region along the south shore of Lake Superior, the Midcontinent rift is structurally defined as a horst bounded by regional reverse fault systems and positioned between wedge-shaped flank basins. This horst, which originated as a graben formed by deep-seated listric, normal fault movements, is typically composed of up to 10,000 m of basalt overlain by units of the Oronto and Bayfield groups. This combination of basalts and sedimentary rocks, extending to depths in excess of 20 km., may represent the greatest thickness of intra-continental rift deposits known on Earth (Behrendt et al., 1988). This anomalously thick "rift-crust", especially as known in the Lake Superior basin, is in contrast to thinner-than-normal crust reported associated with many rift depressions of the world (example, the Baikal rift) (Artemjev and Artyushkov, 1971).

Locally along the central horst of the Lake Superior region one or both of the associated sedimentary rock groups has been entirely eroded, resulting in Pleistocene-age till lying directly on volcanic rock. Formerly, Bayfield strata were believed to unconformably overlie the volcanic basement within the flank basins, but recent drill records from Iowa indicate Oronto rocks were also deposited in these basins (Witzke, 1990).

South of the Lake Superior outcrop region, the MRS is easily traced by gravity and magnetic mapping to its generally accepted extremities in Kansas and Michigan. The gravity signature, a linear central

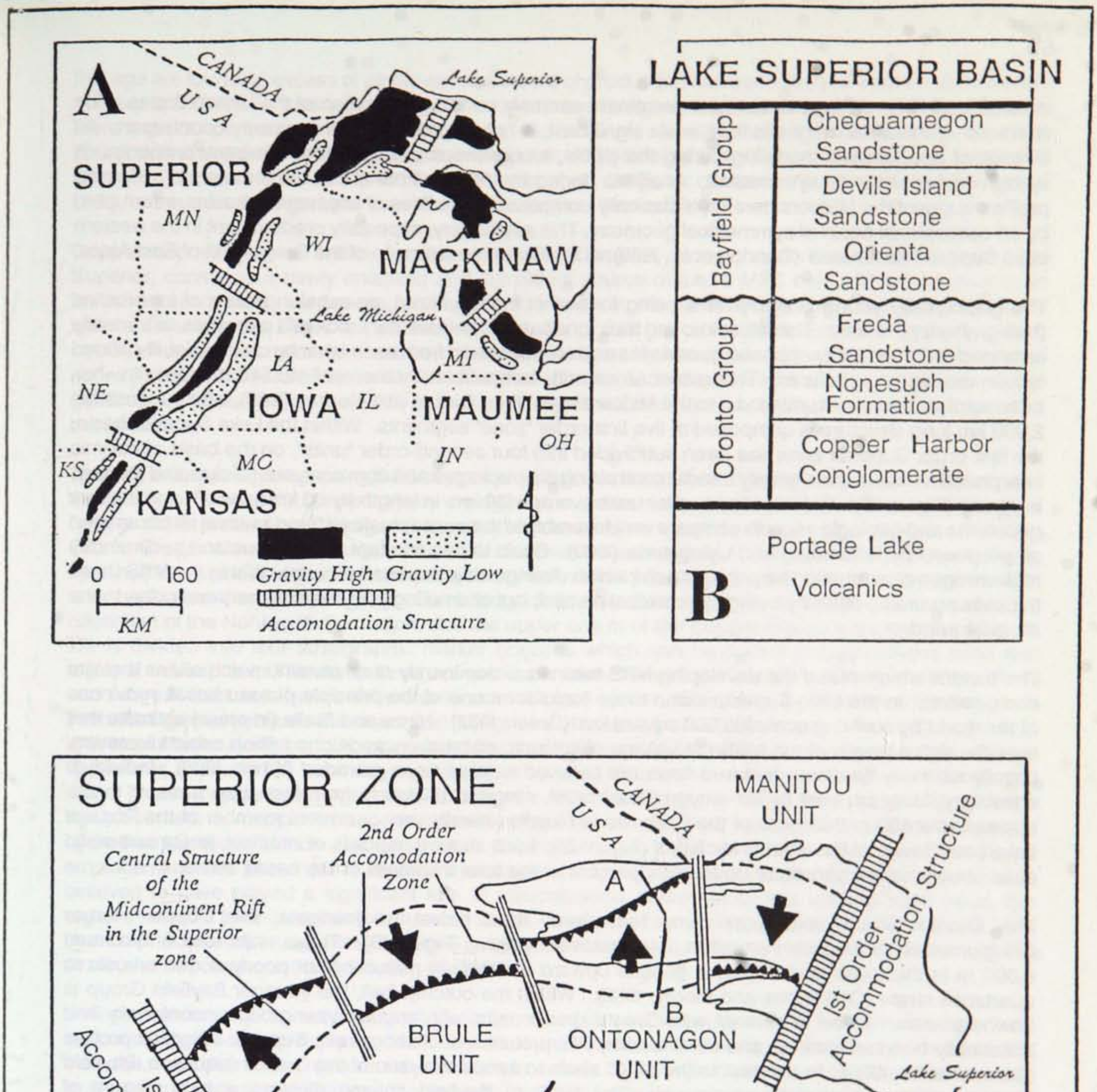


Figure 1. Midcontinent Rift System. A) Regional trend of the Midcontinent Rift System as displayed by its gravity anomaly signature. 1st order zones are shown in bold print. B) Midcontinent rift stratigraphic column as identified in the Lake Superior basin region. C) Differentiation of the Superior Zone (Lake Superior basin) of the Midcontinent Rift System into four structural units. Hachured border of each unit represents active Middle Proterozoic faulting, while dashed border represents the structurally passive flank of the same time period. After Dickas, 1989, and Dickas and Mudrey, 1989.

maxima with flank minima, is the most prominent anomaly on the gravity map of the United States. The character of the MRS magnetic field, while significant, is not as prominent as its gravity counterpart. As a result of geopotential modeling during the 1960's, a consensus emerged that this intracontinental rift system was structurally symmetrical. Analyses during the 1980's of hundreds of kilometers of reflection profiles suggest the Midcontinent rift is basically composed of a series of asymmetric basins, interrupted by an occassional basin of symmetrical geometry. This asymmetry, especially predominant in the western Lake Superior Basin area (Cannon et al., 1989), is similiar to the structure of the Gregory rift of East Africa.

The geologically young Gregory rift is being formed of individualized, en-echelon basins of taphrocline (half-graben) geometry. These basins, and their igneous and sedimentary rock infill packages, are linearly arranged in diametrically opposed geometries and are separated from each other by cross-axial, ill-defined "accommodation" structures. This extant African rift, compatible to other well-studied modern rifts, has been applied as a structural model to the Midcontinent rift by Dickas and Mudrey (1989), who suggest this 2,000 km long structure is composed of five first-order "zone" segments. Within the Lake Superior basin, the first order Superior zone has been subdivided into four second-order "units", on the basis of seismic interpretation, interrupted gravity trends, contrasting dip packages and core analyses (Dickas and Mudrey, in press) (Figure 1C). These second-order units average 150 km. in length by 60 km. in width and in other geometric and geologic aspects compare very favorably to the conceptualized "fundamental rift parameter" as proposed by Rosendahl and Livingstone (1983). Basic to this concept are igneous and sedimentary rock wedges of alternating isopach patterns which distinguish juxtaposed rift units. Within the MRS these thickness patterns differ by being symmetrical parallel, but alternatingly asymmetric perpendicular, to the rift axial trend.

The basalts which infilled the developing MRS basins are dominantly of an aluminum-rich olivine tholeiite composition. In the Lake Superior basin these rocks form one of the principle plateau basalt provinces of the world by covering some 100,000 square km. (Green, 1982). Hinze and Braile (in press) estimate that over the entire length of the MRS the volume of volcanic material exceeds one million cubic kilometers. Locally as many as seven hundred flows are believed to have been extruded (White, 1966). Individual sheets, typically capped by an amygdaloidal crust, range in thickness from less than a meter to the approximate 400 m thickness of the Greenstone Flow, a laterally very persistent member of the Portage Lake Lava Series of the state of Michigan (Figure 1B). Lens shaped deposits of interflow, felsite and mafic clast conglomerate constitute up to eight percent of the total thickness of the basalt series.

The Oronto Group constitutes three formations, from oldest to youngest, the Copper Harbor Conglomerate, Nonesuch Formation, and Freda sandstone (Figure 1B). These units total a maximum 6,000 m in thickness and display a general upward increase in maturity from poorly sorted arkosic to quartzose strata (Ojakangas and Morey, 1982). Within the outcrop belt, the younger Bayfield Group is nowhere known to be in contact with Oronto group units. An angular inter-group unconformity has historically been advocated, and confirmed by interpretation of offshore Lake Superior reflection profiles (Cannon et al., 1989). In contrast to the 60/40 shale to sandstone ratio of the Oronto sequence, Bayfield formations are 99 percent sandstone. The 2,100 m Bayfield column displays a high degree of compositional maturity within the type-section exposures located along the shore of Lake Superior within the state of Wisconsin. South of this outcrop belt Paleozoic era strata cover the MRS.

A principle explanation accounting for the wealth of Lake Superior district geologic and geophysical information relates to extensive analyses of its economic value, both realized (copper) and potential (oil and natural gas). The Lake Superior basalt district is the second greatest copper district in the world after the Bingham, Utah "porphyry copper" area, and from 1844, the initial year of production, until 1887 was the premier copper district of North America. None of the approximate one-hundred mining companies which developed this region are presently in operation.. Historically about 11 billion pounds of copper were produced through 1968 (Weege and Pollack, 1971), including slabs of native copper with an in-situ weight of up to 800 tons. Copper deposits occupy two stratigraphic positions within the Keweenawan basalt sequence composing rift infill. These no longer economic, native-copper ores are found within flow tops and interflow conglomerate strata of the Middle Keweenawan Portage Lake Lava Series. Although mineralization of this type is found to a degree throughout the Lake Superior district wherever basalts of this age are found, in excess of ninety-seven percent of production has been derived from an approximate 200 square kilometer area centered on the Keweenaw Peninsula of the south shore of Lake Superior. More than 90 percent of this production was derived from six mines which extracted ore from three categories of deposits: interflow conglomerate lodes, amygdaloidal lodes, and fissure veins. Copper was emplaced between l,060 and l,047 Ma (Bornhorst et al., l988), as compared to the approximate age of l,095 for the host Portage Lake Lava Series (Paces and Davis, l988).

Copper deposits hosted within the Duluth Gabbro Complex, along the northwestern shore of Lake Superior, constitute a newly analyzed and significant source of future MRS ores. This host-rock is an intrusive equivalent to the basalts forming the basement strata within stratigraphic columns associated with this rift. The ore zones are situated at the base of this gabbro and are formed of copper, nickel, cobalt, and platinum metals derived from the parent magma by interaction with sulfur from the emplaced country rocks. These ores have not been mined to date because of environmental concerns. Other gabbros are affiliated with the MRS, in southern Minnesota and Kansas, but their economic potential is not known.

The only example of present-day MRS copper production is found in the Porcupine Mountain district, near the Michigan-Wisconsin border. Here, a single mine extracts chalcocite (80% production) and native copper (20% production), along with trace amounts of silver, from a strata-bound deposit within the Oronto Group. Although this copper mineralization was known as early as the l850's, the history of mining here has been principally associated with both the American Civil War (initial production began in l865) and the Korean War (present production history dates from l953). The ore-bearing rock, located in the basal 6 m of the Nonesuch Formation, and the upper one m of the Copper Harbor conglomerate (Figure lB), is divided into four stratigraphic marker horizons which can be traced throughout the mine and regional outcrop belt. The mineralized zones are conformable with the bedding of these host members and range in thickness from a millimeter to as much as 6 meters. The principle ore chalcocite occurs as interstitial grains between the sand and silt grains. Native silver, where found, commonly occurs as individual sheets along bedding planes and fractures. White and Wright (l954) believe the origin of these deposits is a function of local degrees of diagenesis and deformation. Fine-grained, i.e., siltstone and shale, hosted copper was largely derived from the waters in which the strata were deposited, whereas the placement of copper in sandstone and conglomerate shows a relation to structure by way of hydrothermal redistribution occurring after lithification and burial. Whatever the specifics of ore emplacement, the presence of organic material in the basal members of the Nonesuch formation is believed to have played a significant role in mineralization. Throughout this underground mine, this organic material is responsible for crude oil seeps emanating from ceiling fractures. These seeps have been a principle impetus for the intensive geophysical evaluation of the hydrocarbon potential of MRS strata and structure conducted during the l980's.

Until several decades ago consensus knowledge stated that sedimentary rock of Precambrian age could not contain hydrocarbon deposits. The absence of pre-Phanerozoic life and reservoir rock characteristics were the most often cited reasons why terranes in excess of 570 million years in age should be avoided by the oil and natural gas industry. Today, with fossil evidence documenting the initiation of life on Earth as early as 3.5 Ga, the discovery in l962 of Precambrian, commercial, indigenous oil and gas reserves within the East Siberian Platform of the Soviet Union (Vassoyevich et al., l97l: Meyerhoff, l980), and the acknowledgement that all the factors necessary for the formation of hydrocarbon deposits have been in operation throughout the past 3.5 billion years, the Precambrian is no longer discounted by the hydrocarbon industry (Dickas, l986a and l986b). Within Precambrian terranes, those representing rifting events have become favored exploration targets because, while rift sedimentary rock accumulations represent 5.5 percent by area of the approximate 600 basins of the world, these rocks account for l0 percent of present world reserves, and may account for as much as 25 percent of hydrocarbons yet to be discovered (Klemme, l980).

The presence of MRS indigenous crude oil in the Upper Peninsula of Michigan, determined by Kelly and Nishioka (l85) to be at least l,047 million years of age through dating of oil-included calcite, spurred exploration programs throughout the full extent of the southwestern area of this rift system during the l980's. With its ten known sedimentary basins, ranging from one to more than l0 km. in depth (gravity calculation), and totaling approximately l50,000 square km., the Midcontinent rift System is being intensely studied through analyses of state-of-the-art geopotential and seismic data. To date, three record-depth wells have been drilled into Midcontinent rift sedimentary rock in Kansas, Iowa, and Michigan, and while all are non-commercial, correlation of their stratigraphic columns has extended the known distribution of Nonesuch Formation source rock at least 750 km. southeast of the Lake Superior oil seep region.

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# COPPER DEPOSITS AND HOST ROCKS, WESTERN UPPER PENINSULA OF MICHIGAN

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An overview of the Midcontinent rift system is provided by Dickas (this volume). An important component of this rift are economic deposits of copper located in the western Upper Peninsula of Michigan. Deposits occur as: native copper or copper sulfide within the Portage Lake Volcanics (PLV) of the Keweenaw Peninsula and copper sulfide and native copper within the Nonesuch Shale at White Pine.

The Portage Lake Volcanics are tholeiitic flood basalts (White, 1960) which host a billion dollar native copper mining district (Figure 1a). Over 200 individual lava flows with a total thickness of 5 km are exposed in the Keweenaw Peninsula. Silicic volcanic and subvolcanic rocks comprise less than 1 volume percent of the PLV. Although interflow sedimentary rocks are less than 5 volume percent (Merk and Jirsa, 1982), they are important ore host rocks. Dikes of mafic and intermediate composition cut the exposed volcanic pile but are as a whole uncommon. These volcanic rocks were erupted over a 2.2+/-1.2 Ma span of time at about 1095 Ma (Davis and Paces, 1990).

Paces (1988) and Paces and Bell (1989) have shown that the composition of the PLV is different than other major continental flood basalts in that the PLV contains abundant magnesia-rich, high alumina olivine tholeiites which are relatively primitive. Geochemical stratigraphy within the basalts is cyclical with minor and major cycles and a general trend, all a result of igneous processes (Paces, 1988). The basaltic magmas were derived by partial melting of relatively shallow, sub-continental upper mantle with the younger basalts being more primitive and less contaminated by crustal material (Paces and Bell, 1989). Model calculations show that major cycles are due to complex fractional crystallization and replenishment in large magma chambers near the crust/mantle interface (Paces, 1988). Minor cycles and silicic rocks are a result of closed system fractional crystallization in small magma chambers within the crust. Degassing of volatiles, particularly $SO_2$, during and after eruption in an oxidizing subaerial environment, was important because it created sulfur-deficient lava flows which favored the later deposition of native copper.

The typical subaerial lava flow consists of a thin vesicular base overlain by a massive (vesicle-free) interior capped by flow top. There are three main varieties of flow top recognized in the PLV: 1) flow top breccia; 2) vesicular; and 3) flow top breccia with a sandy or silty matrix. White (1968) estimated that 21 percent of the lava flow tops in the PLV are brecciated. The uppermost 5 to 20 percent of most individual lava flows are vesicular with between 5 and 50 percent vesicles which are commonly filled with secondary minerals. Since the fundamental control on the movement of ore fluids is per-

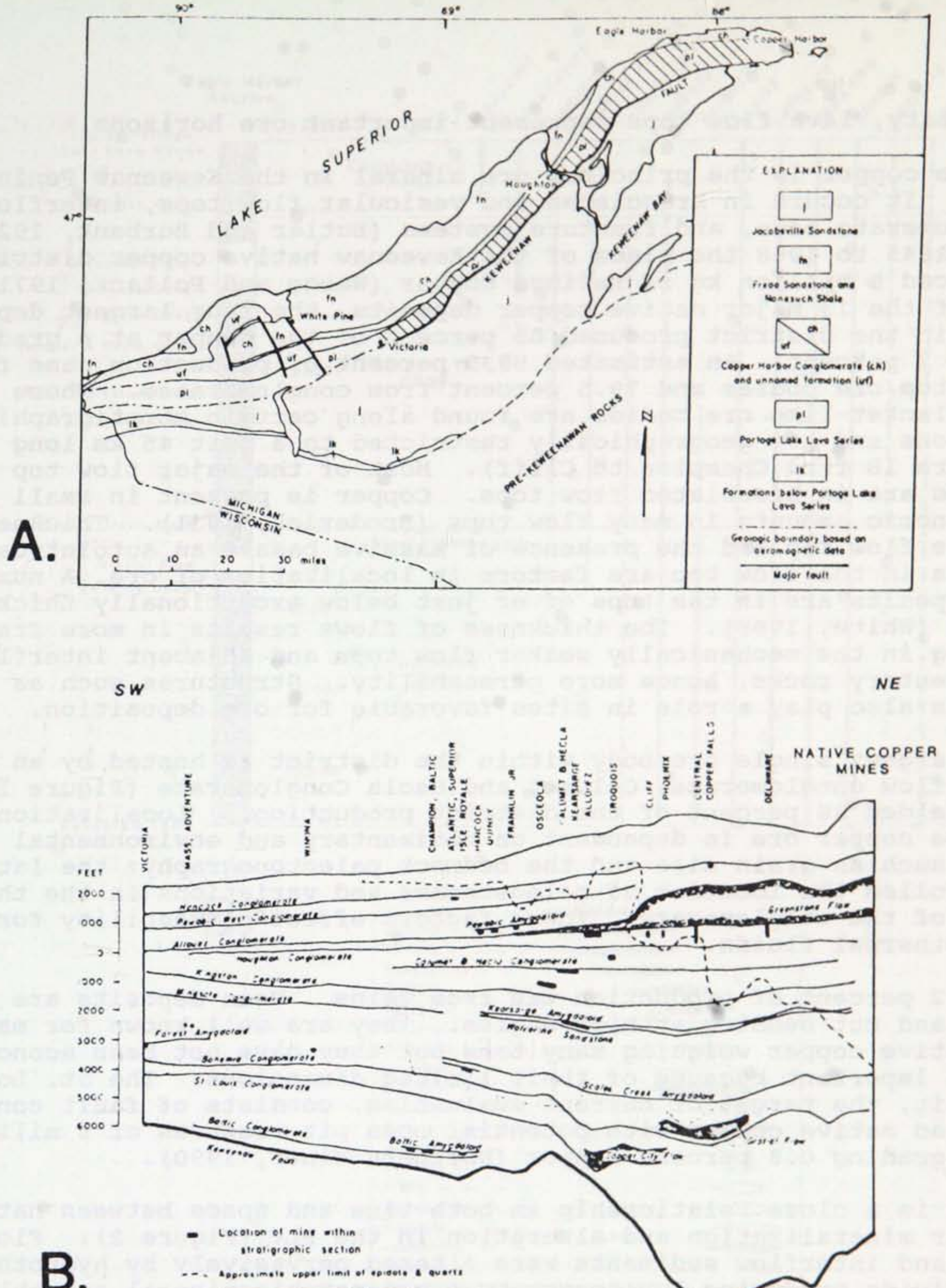


Figure 1: A. Generalized geologic map of upper Precambrian rocks of western Upper Peninsula, Michigan. Hatched area is represented in cross-section in B. B. Generalized stratigraphic section of the Portage Lake Volcanics from Victoria to Copper Harbor (modified from Stoiber and Davidson, 1959). The major marker horizons and mines are shown. The dashed and dotted lines represent the approximate stratigraphic limits of secondary epidote and quartz and prehnite respectively.

meability, lava flow tops represent important ore horizons.

Native copper is the principal ore mineral in the Keweenaw Peninsula. It occurs in brecciated and vesicular flow tops, interflow conglomerate beds, and fracture systems (Butler and Burbank, 1929). From 1845 to 1968 the mines of the Keweenaw native copper district produced 5 billion kg of refined copper (Weege and Pollack, 1971). Out of the 13 major native copper deposits, the four largest deposits in the district produced 85 percent of the copper at a grade of about 2 percent. An estimated 58.5 percent of production came from flow top ore bodies and 39.5 percent from conglomerates. These lens and blanket-like ore bodies are found along certain stratigraphic horizons and are geographically restricted to a belt 45 km long (Figure 1B from Champion to Cliff). Most of the major flow top ore bodies are in brecciated flow tops. Copper is present in small uneconomic amounts in many flow tops (Broderick, 1931). Thickness of the flow top and the presence of massive basalt as autointrusive bodies in the flow top are factors in localization of ore. A number of deposits are in the tops of or just below exceptionally thick flows (White, 1968). The thickness of flows results in more fracturing in the mechanically weaker flow tops and adjacent interflow sedimentary rocks, hence more permeability. Structures such as faults also play a role in sites favorable for ore deposition.

The largest single ore body within the district is hosted by an interflow conglomerate, Calumet and Hecla Conglomerate (Figure 1B). It yielded 38 percent of the district production. Localization of native copper ore is dependent on sedimentary and environmental factors such as grain size and the bedrock paleotopography; the latter controlled the location of paleostreams and variations in the thickness of the conglomerate. These factors effect permeability for hydrothermal fluids.

Only 2 percent of production was from veins. Vein deposits are tabular and cut bedding at high angles. They are well known for masses of native copper weighing many tons but they have not been economically important because of their limited dimensions. The St. Louis deposit, the target of current evaluation, consists of fault controlled native copper with potential open pit reserves of 8 million tons grading 0.8 percent copper (Northern Miner, 1990).

There is a close relationship in both time and space between native copper mineralization and alteration in the PLV (Figure 2). Flow tops and interflow sediments were altered pervasively by hydrothermal fluids producing low-temperature metamorphic mineral assemblages (Stoiber and Davidson, 1959; Jolly and Smith, 1972). These minerals occur as amygdule and vein fillings and as whole rock replacements in the most permeable units. Intensity and degree of alteration varies as a function of position within individual flows, position in the volcanic pile and proximity to cross-cutting fractures. Metamorphic zoning varies vertically within the volcanic pile and is equivalent to zeolite and prehnite-pumpellyite facies (Stoiber and Davidson, 1959; Livnat, 1983). Copper deposits tend to be within the pumpellyite zone.

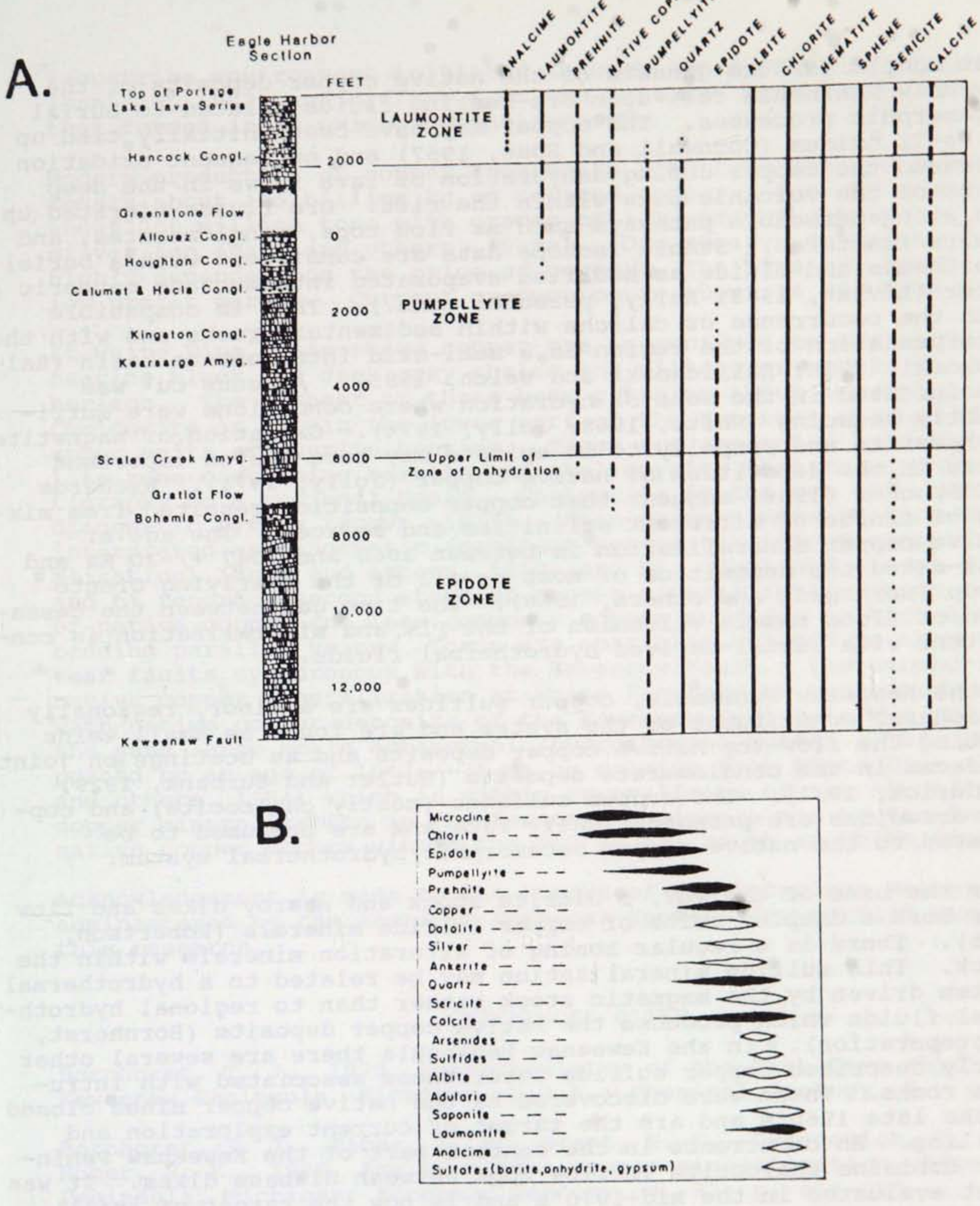


Figure 2: A. Distribution of secondary minerals in the Eagle Harbor section of the Portage Lake Volcanics (compiled from Butler and Burbank, 1929; Jolly, 1974; Jolly and Smith, 1972; Stoiber and Davidson, 1959; White, 1968). Location of section is between Copper Falls and Delaware Mines shown in Figure 4. B. Paragenesis of secondary minerals in the flow tops and veins (from White, 1968). Solid black symbols are the more abundant minerals. Secondary minerals shown here are nonmagmatic and not of supergene origin.

Most models for the genesis of the native copper deposits of the Keweenaw Peninsula call upon ore-bearing fluids related to burial metamorphic processes. The copper may have been initially tied up in Fe-Ti oxides (Cornwall and Rose, 1957) and subsequent oxidation released the copper during dehydration of lava flows in the deep parts of the volcanic pile within the rift. Ore fluids migrated up dip along permeable pathways such as flow tops, conglomerates, and faults/fractures. Stable isotope data are consistent with a burial hypothesis and fluids as modified evaporated intermontane meteoric water (Livnat, 1983; Kelly, personal comm.). This is compatible with the occurrence of caliche within sedimentary rocks and with the interpretation of the region as a semi-arid intermontane basin (Kalliokoski, 1986; Kalliokoski and Welch, 1985). Aqueous $Cu^+$ was precipitated in the zone of hydration where conditions were sufficiently reducing (White, 1968; Jolly, 1974). Oxidation of magnetite to hematite and pumpellyite to epidote may have played important roles in the deposition of native copper (Jolly, 1974). Richards and Spooner (1986) suggest that copper deposition resulted from mixing of fluids of different salinities and sources. The age of native copper mineralization is between 1060 and 1047 +/-20 Ma and post-dated the deposition of most or all of the overlying Oronto Group (Bornhorst and others, 1988). The time gap between the cessation of flood basalt volcanism of the PLV and mineralization is consistent with burial-derived hydrothermal fluids.

In the Keweenaw Peninsula, copper sulfides are a minor, regionally peripheral constituent of the system and are found as small veins cutting the flow-top native copper deposits and as coatings on joint surfaces in the conglomerate deposits (Butler and Burbank, 1929; Broderick, 1931). The copper sulfides (mostly chalcocite) and copper arsenides are paragenetically late and are presumed to be related to the native copper metamorphic/hydrothermal system.

Near the base of the PLV, a diorite stock and nearby dikes and flow tops host a complex suite of copper sulfide minerals (Robertson, 1975). There is a regular zoning of alteration minerals within the stock. This sulfide mineralization may be related to a hydrothermal system driven by the magmatic stock rather than to regional hydrothermal fluids which produced the native copper deposits (Bornhorst, in preparation). In the Keweenaw Peninsula there are several other poorly described copper sulfide occurrences associated with intrusive rocks. These were discovered as the native copper mines closed in the late 1960's and are the target of current exploration and drilling. An occurrence in the central part of the Keweenaw Peninsula contains chalcocite in flow tops between diabase dikes. It was first evaluated in the mid-1970's and is now the target of infill drilling. Probable reserves are 3.1 million tons grading 2.95 percent copper (Northern Miner, 1990).

Copper sulfides and native copper in economic quantities are hosted by the Nonesuch Shale at the White Pine Mine (White and Wright, 1954 and 1966; Ensign and others, 1968; Brown, 1971; White, 1971). The Nonesuch Shale is a succession of gray to black siltstone, shale and sandstone within the Oronto Group which overlies the PLV (see Dickas, this volume). These sediments were deposited in a reducing, lacustrine environment initiated through the disruption of drainages (Daniels, 1982) and differs from the over and underlying redbeds that formed in an oxidizing environment.

Modern production of copper from White Pine from 1953 to present totals about 1.5 billion kg of refined copper. Current reserves are about 200 million tons with grades of 1.1 percent copper and 9 grams silver/ton (Mauk and others, 1989a). Ore reserve estimates are highly dependent on the price of copper. The mine has operated on a low profit margin. Current production is about 14,000 tons/day.

At White Pine, main-stage copper ore is confined to chalcocite-bearing black and dark gray shales and siltstones within the mined horizon. The richest of these beds contains about 3 percent copper. Chalcocite is within the lower part of the Nonesuch whereas the upper units are pyrite bearing. These are separated by a blanket-like zone containing bornite and chalcopyrite that cross-cuts stratigraphy on a regional basis. Chalcocite in the ore zone replaces diagenetic pyrite (Brown, 1971). Main-stage mineralization is interpreted as diagenetic (sediment-hosted stratiform copper mineralization) (Mauk and others, 1989a and b). Mauk and others (1989a and b) report a second stage of copper mineralization which consists of native copper and less commonly copper sulfides in high-angle and bedding parallel veins. These are spatially related to thrust and tear faults synchronous with the Keweenaw Fault. The second-stage native copper mineralization at White Pine may be contemporaneous with native copper deposits of the Keweenaw Peninsula and related to the same large scale metamorphic/hydrothermal system. This is supported by an age of 1047+/-35 Ma on calcite from White Pine (Ruiz and others, 1984) which is within error limits of the age of native copper mineralization in the Keweenaw Peninsula and by the fact that native copper occurs within the PLV as far south as White Pine.

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A REVIEW OF THE ECONOMIC SIGNIFICANCE OF THE DULUTH COMPLEX, NE MINNESOTA

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The Duluth Complex consists of dominantly mafic igneous rocks of Keweenawan Age (1.1 Ga) that are exposed in an arcuate body extending from Duluth to Hovland, MN. These rocks are divided into an older anorthositic series and a younger troctolitic series (Figure 1). Past exploration of basal troctolitic rocks has outlined five kinds of ore mineral occurrences: 1. Cu-Ni sulfides that occur either at the base of the gabbro (basal) or about 300 m above the base (cloud); 2. late stage Pd-Cu-Au enrichment zones within the Cu-Ni mineral deposits; 3. Pt-Pd mineralization related to Cr spinels within layered rocks at the very base of the Complex; 4. Ti-Fe-V-rich oxide ultramafic bodies that intrude the troctolites; and 5. hybrid deposits that may be mixtures of 1 through 4 (for example, the Water Hen intrusion, Figure 2).

The basal Cu-Ni sulfides are estimated to contain greater than 4.4 billion tons of 0.66% Cu with a Cu to Ni ratio of 3.3:1 (Listerud and Meineke, 1977). These sulfides generally occur up to 100 meters above the footwall contact of the Complex in troctolites and gabbros, which intrude country rocks that vary from granite of the Giants Range Batholith in the north, through Biwabik iron formation in the central part, to slates and greywackes of the Virginia Formation in the south (Figure 2). Sulfur and oxygen isotope studies as well as S/Se ratios show that the sulfur source appears to come from pyrite and pyrrhotite rich sedimentary rocks of the Virginia Formation (Ripley, 1981, 1990a,b; Ripley and Al-Jassar, 1987; Rao and Ripley, 1983 for example). A schematic cross-section through the Dunka Road deposit from Severson (1988) is shown in Figure 3.

These sulfide occurrences contain very large resources of platinum group elements (PGE) (Morton and Hauck, 1987,1989). Weighted averages for combined Pt and Pd values for some of the occurrences vary from 378 ppb at Minnamax to 651 ppb at Spruce Road. Some zones within the Cu-Ni occurrences are enriched in Pd, Pt, and Au (1.5 to 10 ppm), such as in the South Filson Creek and Dunka Road deposits. These enriched zones appear to be related to late stage shear and alteration zones that have been documented in both silicate and sulfide mineralogy. Anomalous PGE areas are related to a period of secondary Cu enrichment. Figure 4 is a schematic drawing of the South Filson Creek Cu-Ni occurrence (Kuhns et al., 1990) that shows the relationship between faulting, silicate alteration and PGE enrichment.

The third type of mineralization occurs within the Birch Lake area (drill hole DU-15). A 3 meter oxide layer (magnetite-ilmenite-Cr-spinel-olivine) occurs at the base of the Complex at the contact with Archean granite footwall (Figure 2 and 5). This layer contains a combined PGE content of 5099 ppb (Sabelin, 1986). Similar zones have been identified in DU-9, located 800 m to the southeast (Dahlberg et al, 1989).

The Ti-Fe-V-rich oxide bodies include the Longnose peridotite, the Longear pyroxenite, and Section 17 oxide bodies that all lie on a northeast trend interpreted to be a pre-Duluth Complex fault (Severson, 1988). Other Ti-Fe oxide-rich bodies have been mapped and/or drilled in the Skibo area, the Boulder Lakes area (Bonnichsen, 1972) and in the Fish Lake area (the latter two are not shown on Figure 2). For example, the Longnose peridotite consists of a small funnel-shaped (150 x 800 m in plan) body (Figures 6, 7, and 8) that is zoned from clinopyroxenite on the edge to oxide-rich dunite in the core (Linscheid, 1991). Massive oxide lenses found in the dunitic core are composed of ilmenite and V-rich (up to 2.5 wt% $V_2O_3$) titanomagnetite. This deposit has reserves of 20 to 50 million tons of 20-25% $TiO_2$ (Bill Ulland; July, 1990).

The Water Hen intrusion (Figure 2, 9 and 10) is a good example of a hybrid Cu-Ni and Ti deposit. It is composed of oxide-rich dunite and peridotite which intrudes into the basal troctolites, some of which are cyclically layered (Mainwaring and Naldrett, 1977; Strommer et al., 1990). The cyclically layered troctolite contains disseminated and massive sulfides and as well, the oxide-rich peridotite/dunite contains massive pyrrhotite with minor Cu and Ni-sulfides at its base. There are also zones of PGE enrichment within the Water Hen intrusion (Morton and Hauck, 1987, 1989 and

Morton, 1989). The ultramafic rocks contain variable amounts of ilmenite and titanomagnetite with reserves of 50 to 70 million tons of 15% $TiO_2$ (Bill Ulland; July, 1990). Footwall rocks in the area are both Virginia Formation and meta-basalts. The intrusion contains inclusions of both.


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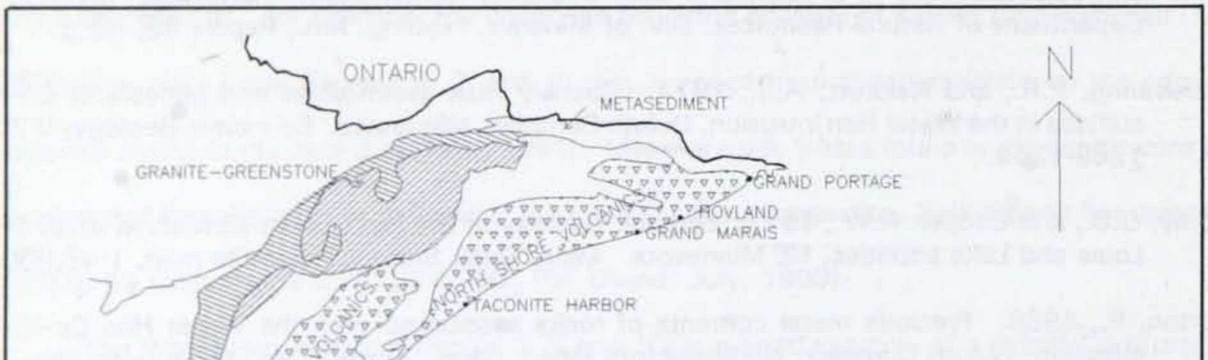


Figure 1. Geology of the Duluth Complex

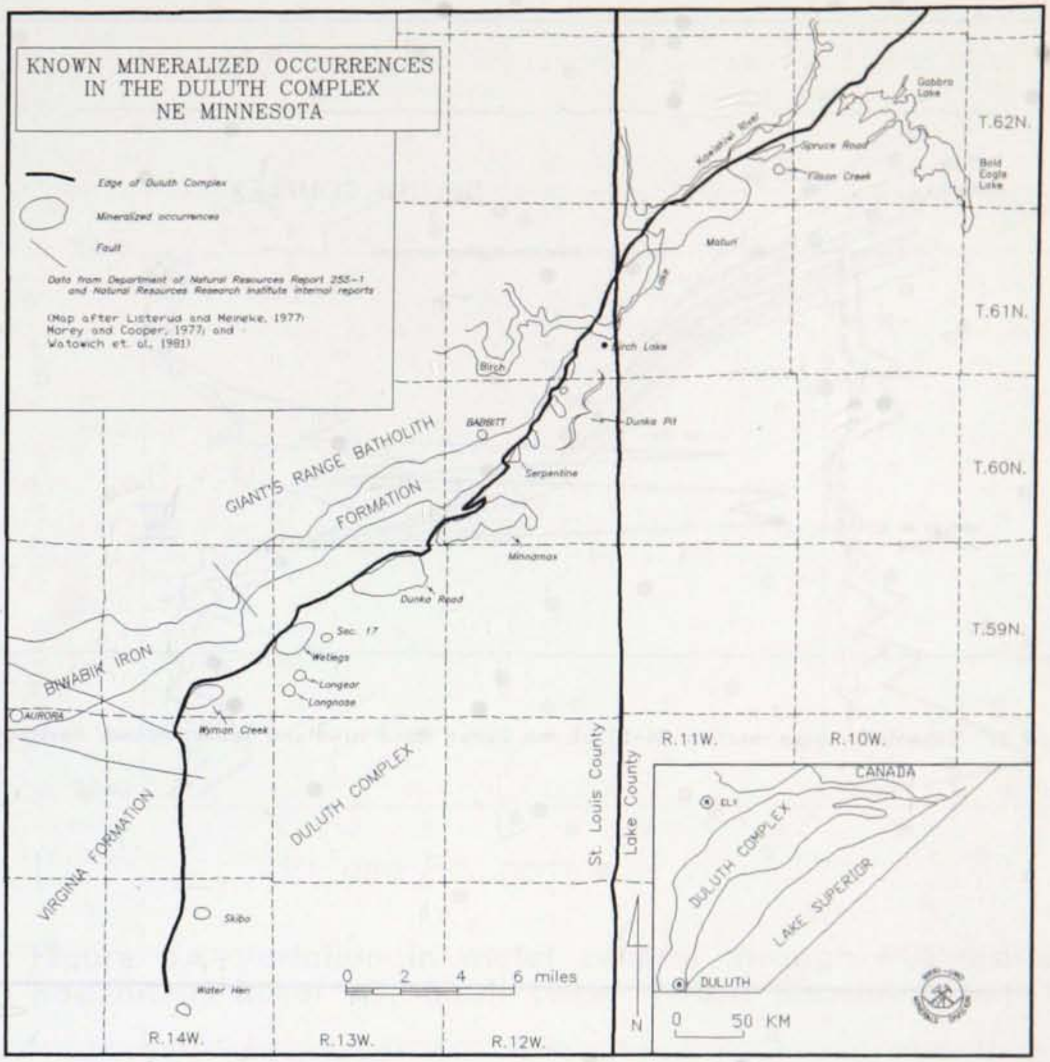


Figure 2. Location of Cu–Ni and Fe–Ti deposits within the Duluth Complex

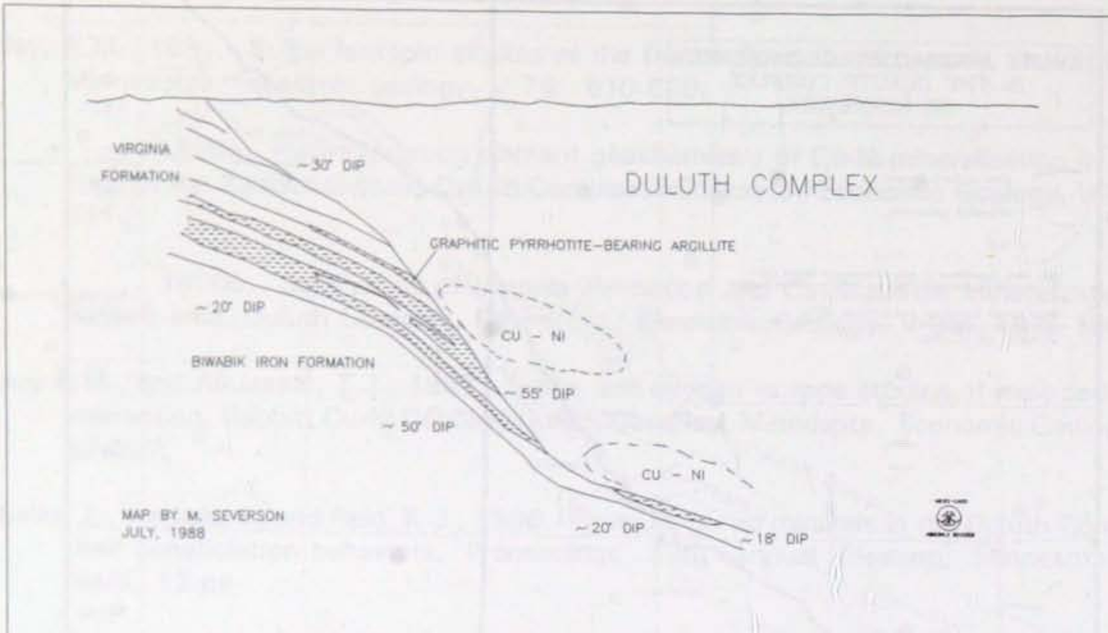


Figure 3. Schematic cross-section (N-S) of the Dunka Road area (no scale implied; Severson, 1988)

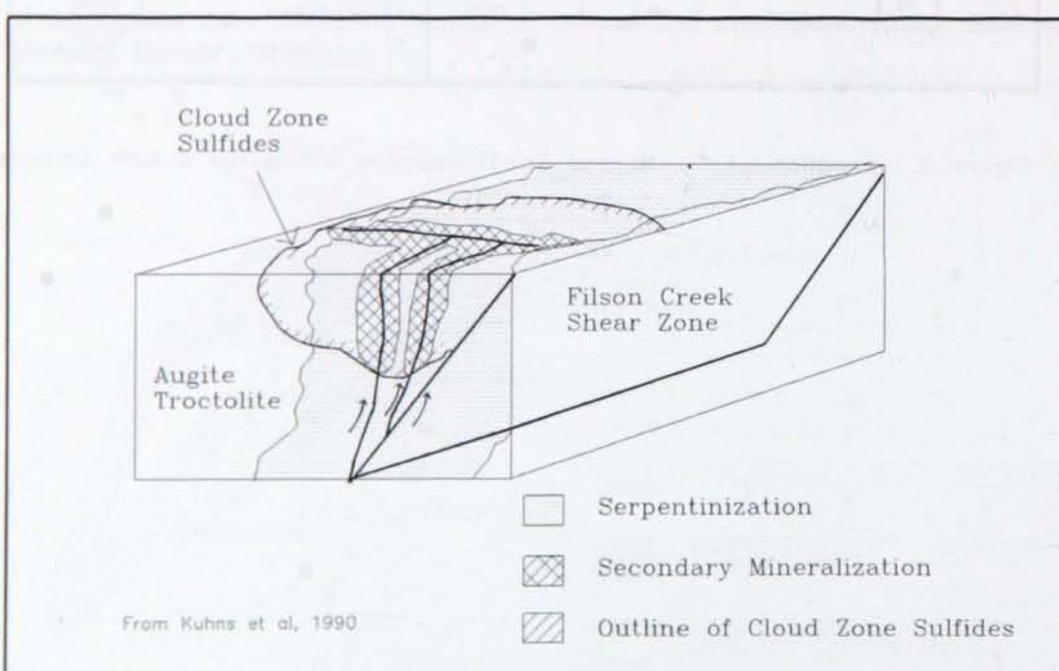


Figure 4. Schematic drawing of South Filson Creek Cu-Ni showing

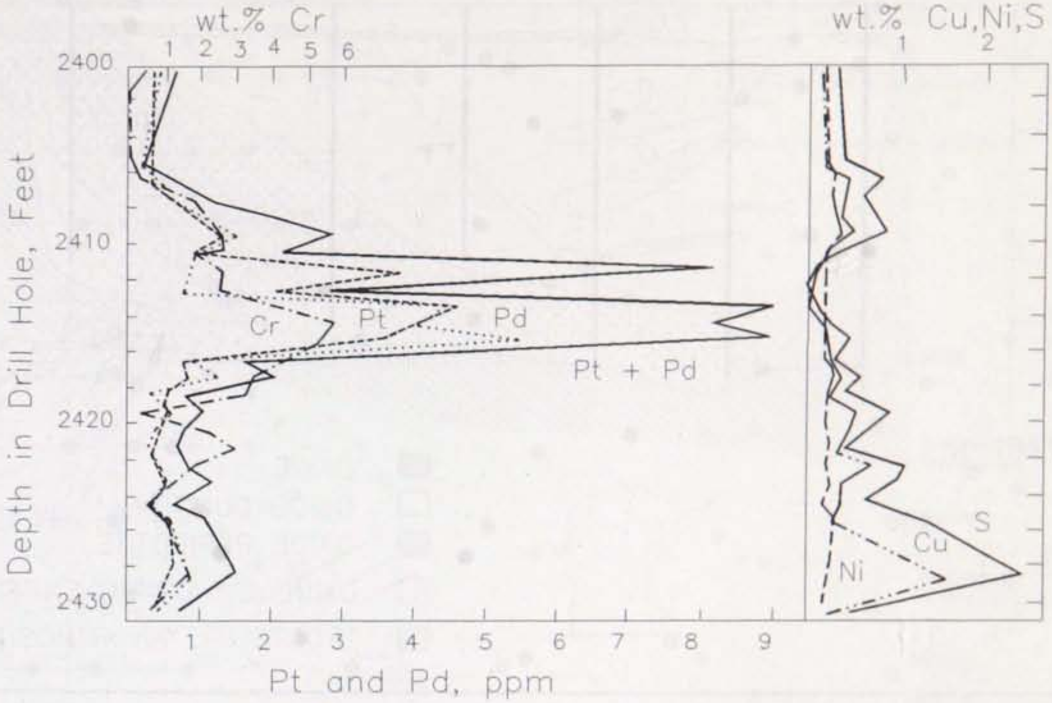


Figure 5. Variation in metal content through PGE–enriched horizon in Duval 15, Birch Lake (from Sabelin et al, 1986)

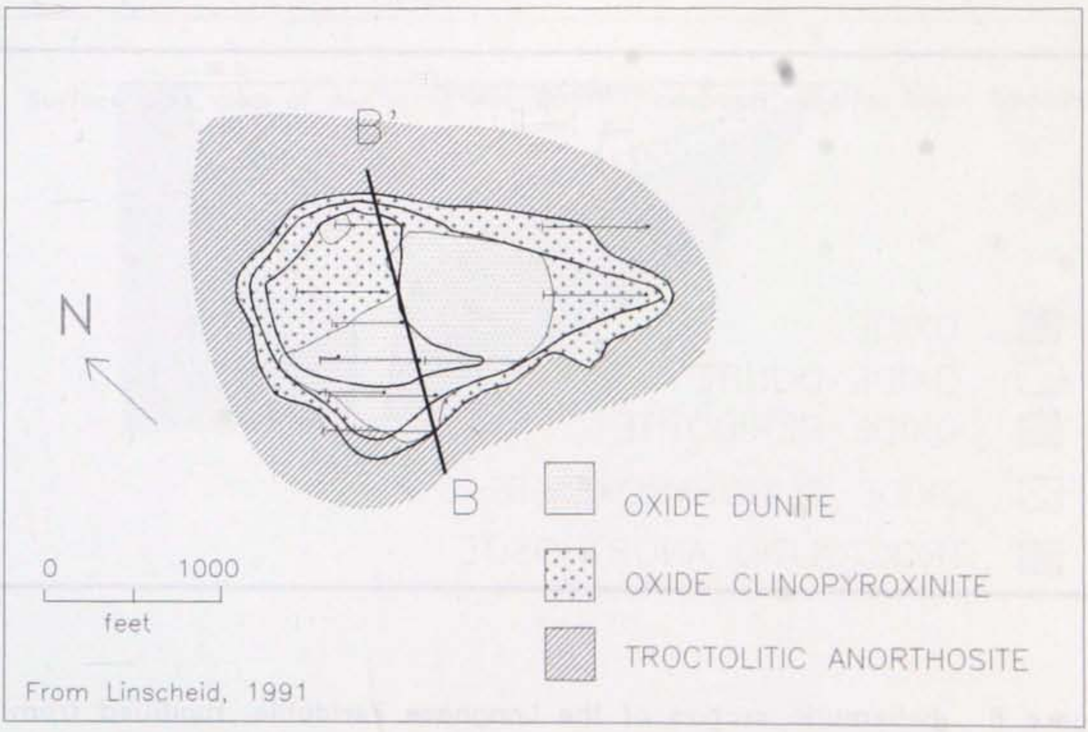


Figure 6. Gravity of the Longnose Peridotite. Solid black lines are gravity contours, numbers are confidential. B–B' is position of Figure 7.

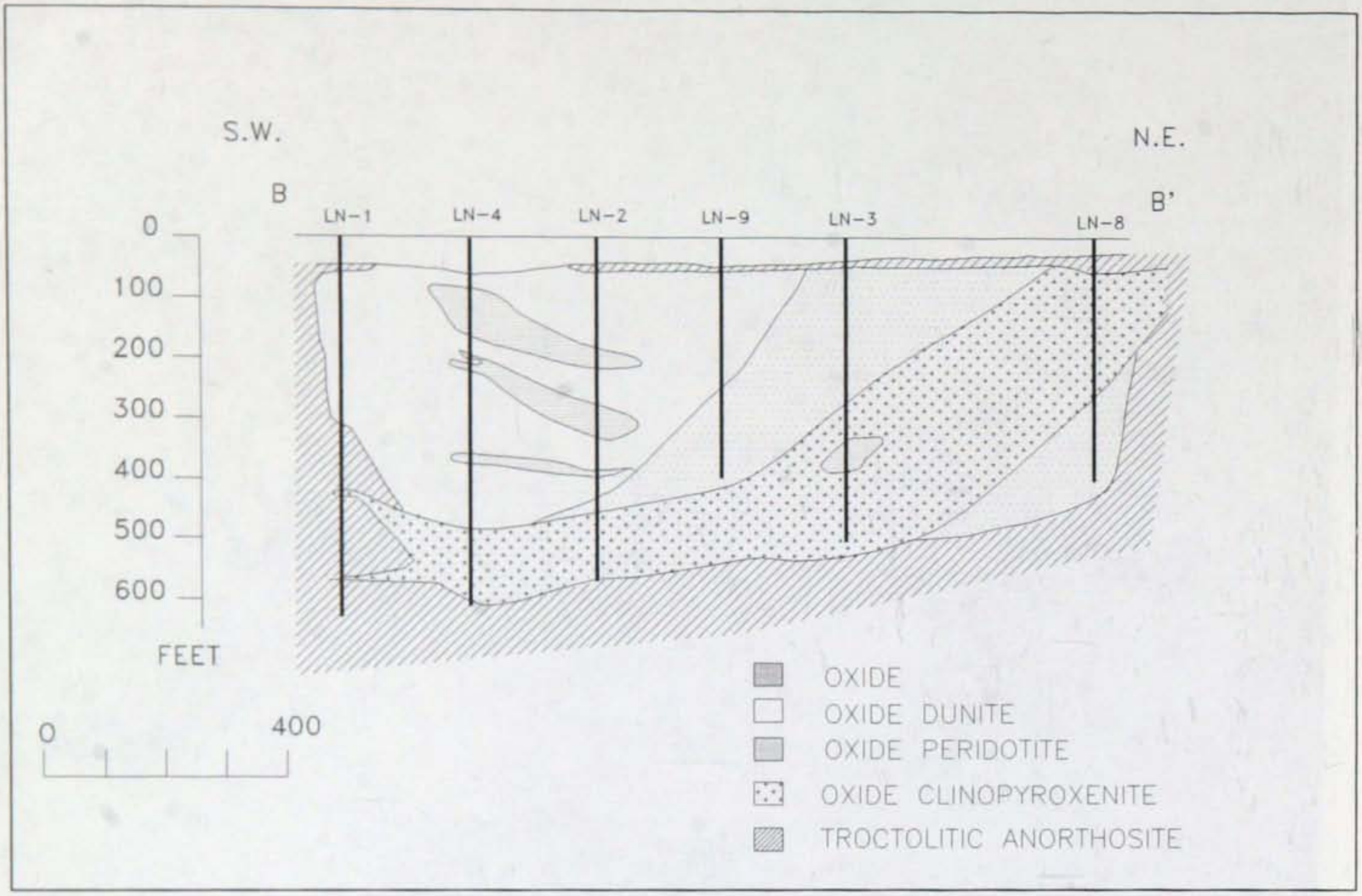


Figure 7. Cross-section through Longnose Peridotite. The cross-section is in the plane of the drill holes that dip 45 degrees NW (modified from Linscheid, 1991)

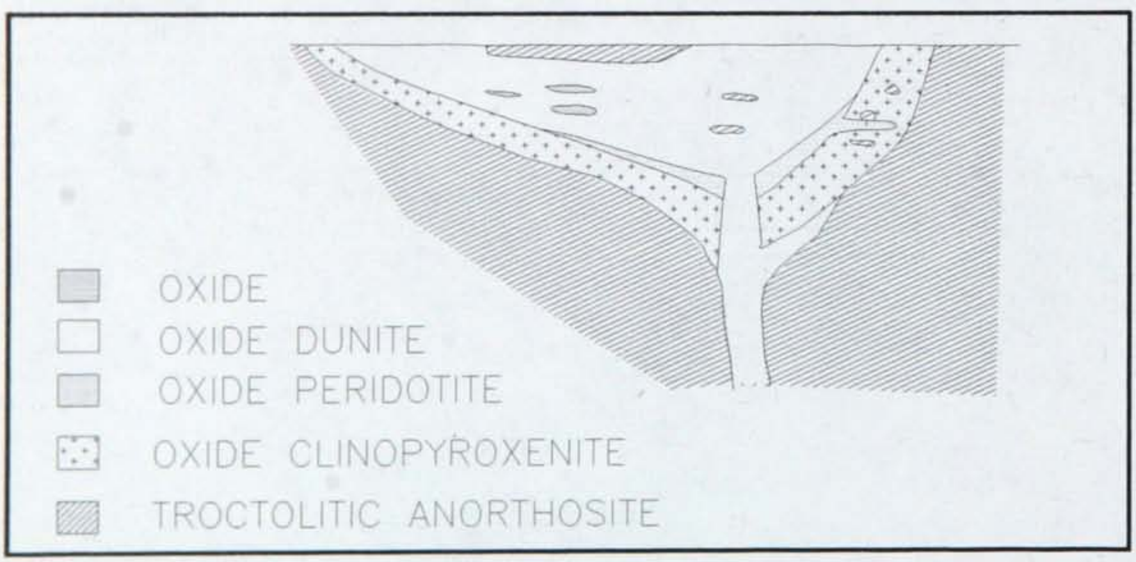


Figure 8. Schematic section of the Longnose Peridotite, modified from Linscheid, 1991

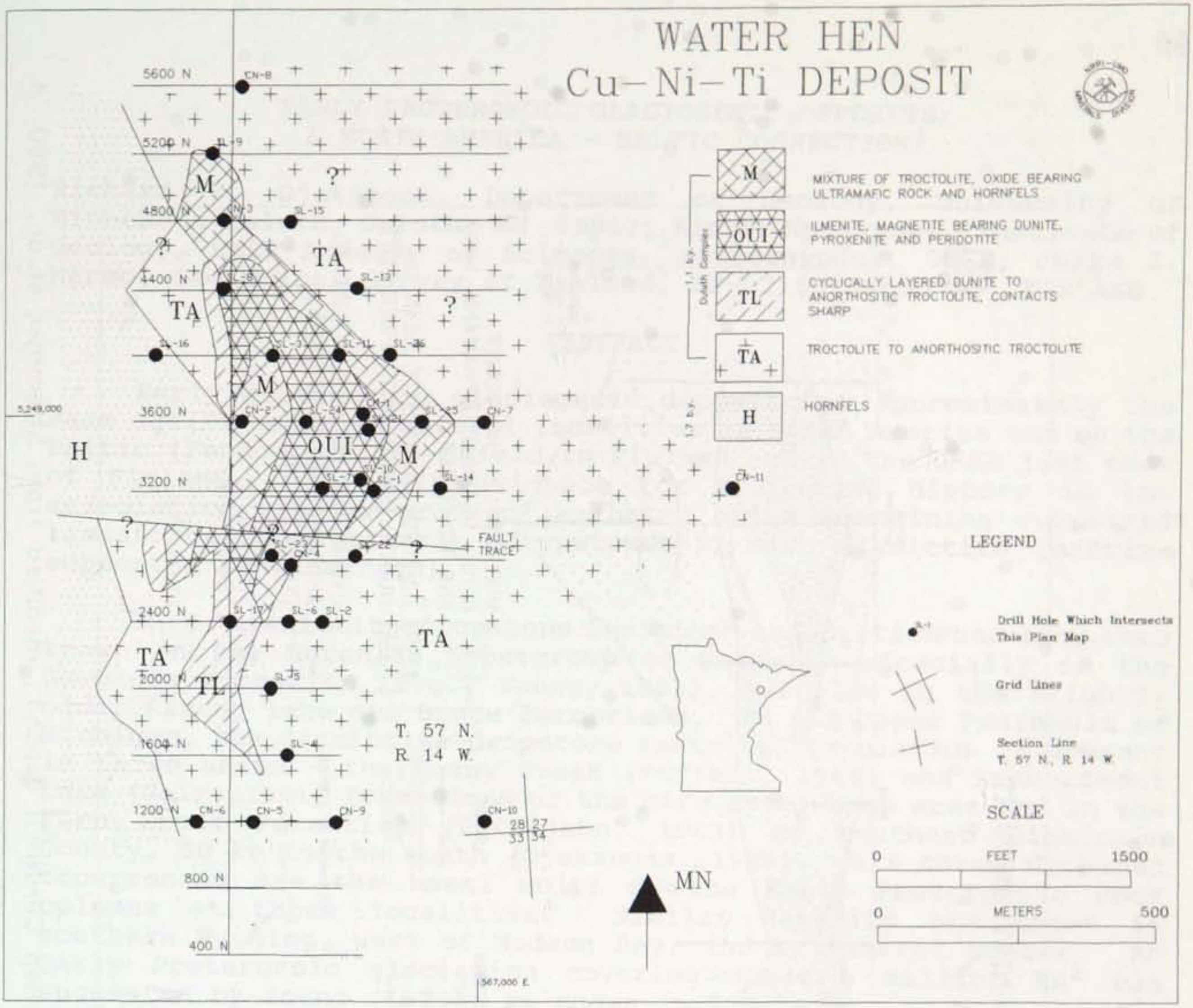


Figure 9. Surface plan map of the Water Hen Cu-Ni-Ti deposit, modifed from Strommer et al., 1991

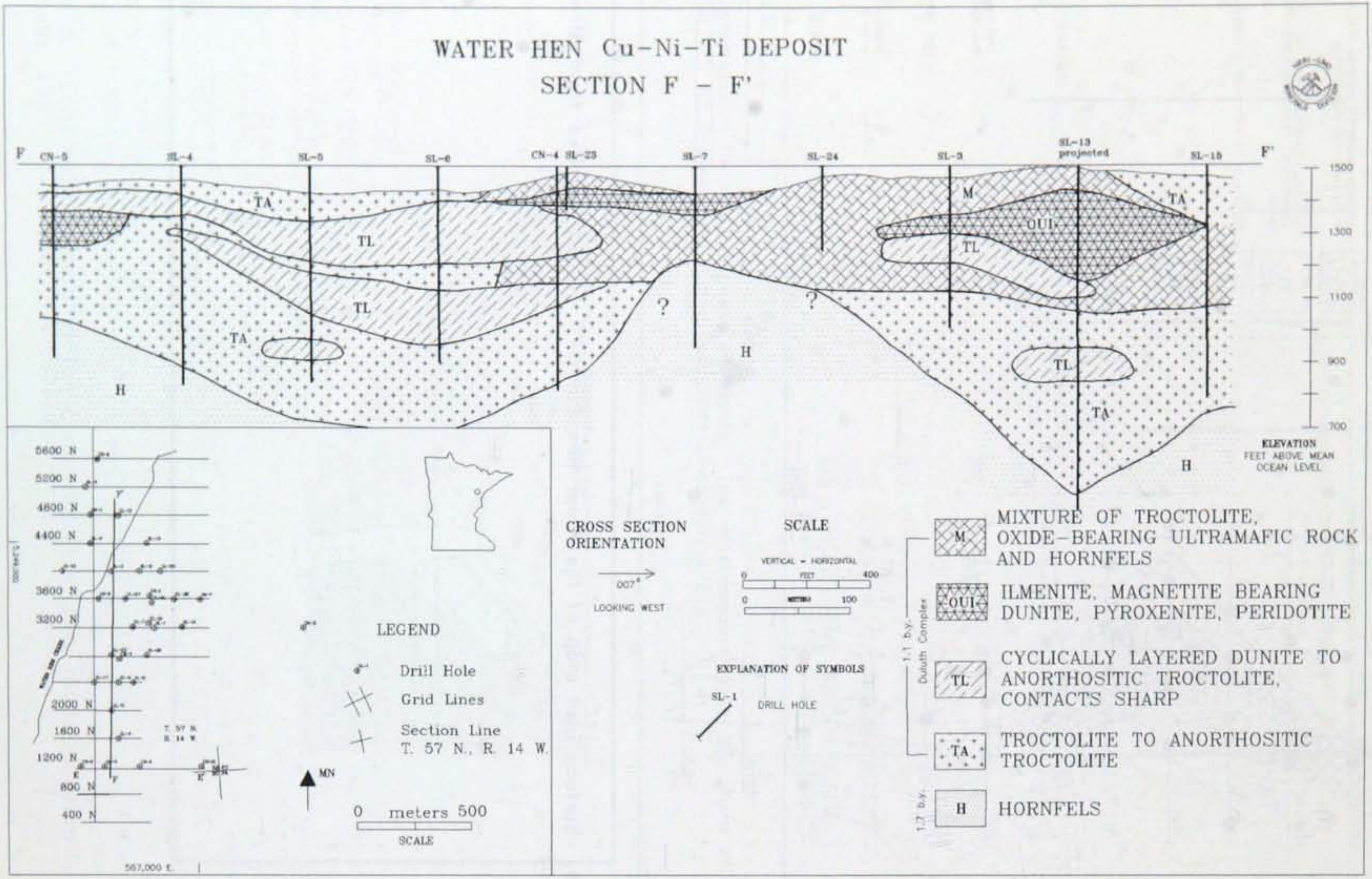


Figure 10. Longditudinal section of the Water Hen Cu-Ni-Ti deposit , modified from Strommer et al., 1990

# EARLY PROTEROZOIC GLACIOGENIC DEPOSITS: A NORTH AMERICA - BALTIC CONNECTION?

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ABSTRACT

Early Proterozoic glaciogenic deposits of approximately the same age(?) occur at several localities in North America and on the Baltic (Fennoscandian) Shield in Finland and in the USSR just east of Finland. The main evidence for a glacial history is the association of fine-grained laminated units containing oversized limestones (some clearly "dropstones"), with diamictite (matrix-supported conglomerate).

This "diamictite-dropstone laminite" association has long been known in the Huronian Supergroup of Ontario, especially in the Gowganda Formation (e.g., Young, 1981), but also in the slightly older Ramsey lake and Bruce Formations. In the Upper Peninsula of Michigan, the diamictite-dropstone laminite association is present in three units - the Reany Creek (Puffett, 1969) and Enchantment Lake (Gair, 1981) Formations of the Marquette Range area and in the Fern Creek Formation (Pettijohn, 1943) of southern Dickinson County, 80 km to the south (Ojakangas, 1984). All three Michigan occurrences are the basal units of the Early Proterozoic rock columns at those localities. Similar deposits are known in southern Wyoming, west of Hudson Bay, and in central Quebec. An Early Proterozoic glaciation covering about 5 million $km^2$ was suggested by Young (1970), as shown in Figure 1.

Such deposits were recently discovered in eastern Finland in the Sariolian Group of the Karelian Supergroup (Marmo and Ojakangas, 1984). The Urkkavaara Formation, which we have interpreted as glaciomarine, is subdivided into four informal members as follows: a lower argillite member, a graded sandstone member, an upper siltstone-argillite member, and a diamictite member. The lower three members contain dropstones, and the upper siltstone-argillite grades upward into the diamictite facies. Subsequently, Marmo (1986) described three additional members that overlie the initial sequence as follows: an upper graded sandstone member, a parallel-bedded conglomerate member and a massive conglomerate member. He interpreted these three members to be comprised of glaciofluvial sediments. The Urkavaara Formation, about 300 m thick, is capped by a 100 m thick meta-regolith that appears to have been largely conglomeratic rock prior to intensive Early Proterozoic weathering. Close inspection of the same stratigraphic horizon in the Sariolian Group has resulted in the recognition of glaciogenic characteristics in five other areas in eastern and central Finland; some of these localities are separated by 350 km (Fig. 2).

Negrutsa and Negrutsa (1981a,b) described 15 diamictite localities in the Sariolian Group and its equivalents, many with associated lonestones, in Karelia, USSR; they proposed sedimentary-tectonic and volcanotectonic origins. We visited seven localities in 1988, and diamictite and lonestone units are associated in the rock column at each. At one locality, River Luzhma, a thick sequence of diamictite is overlain by thinly laminated metasiltstone with dropstones that clearly penetrate and deform the laminae (Ojakangas et al, 1989); interestingly, Eskola suggested a glacial origin for this diamictite in 1917.

The presence of glaciogenic lithologies in widely spaced areas of sedimentary rocks on the Baltic Shield allows for correlation, as a glaciation is an uncommon "mega-event" related to climatic change (Ojakangas, 1988). If all of these deposits are indeed of the same age, the area on the Baltic Shield that was affected by this Early Proterozoic glaciation had a minimum size of about 500 km by 250 km (Fig. 2). If the other diamictite (tilloid) occurrences described by Negrutsa and Negrutsa on the Kola Peninsula in northwesternmost USSR are included, although we have been unable to study those localities to attempt to verify glacial characteristics, the area directly affected by the Early Proterozoic glaciation may be on the order of 200,000 $km^2$, suggestive of a continental-scale glaciation.

The rock units in which these glaciogenic rocks occur in Finland and East Karelia are underlain by mafic lava flows dated at about 2450 m.y. and are intruded by dikes and sills dated at about 2180 to 2160 m.y. (Jatulian diabases). The Huronian Supergroup has 2450 m.y. mafic lavas at its base and is intruded by Nipissing dikes and sills dated at about 2180 to 2150 m.y. These relationships are shown in Figure 3. A contemporaneous or penecontemporaneous glaciation seems likely, and it is possible that the North American and Baltic Shields were in close proximity approximately 2300-2200 m.y. ago (Ojakangas, 1988), as depicted in Figure 4.

The glaciogenic deposits on these two shields appear to be unique lithologies which, when coupled with other data, should be useful in more detailed comparisons of the geologic history of Early Proterozoic time. Certainly they show that cooperative projects by geologists working on both shields can be productive.

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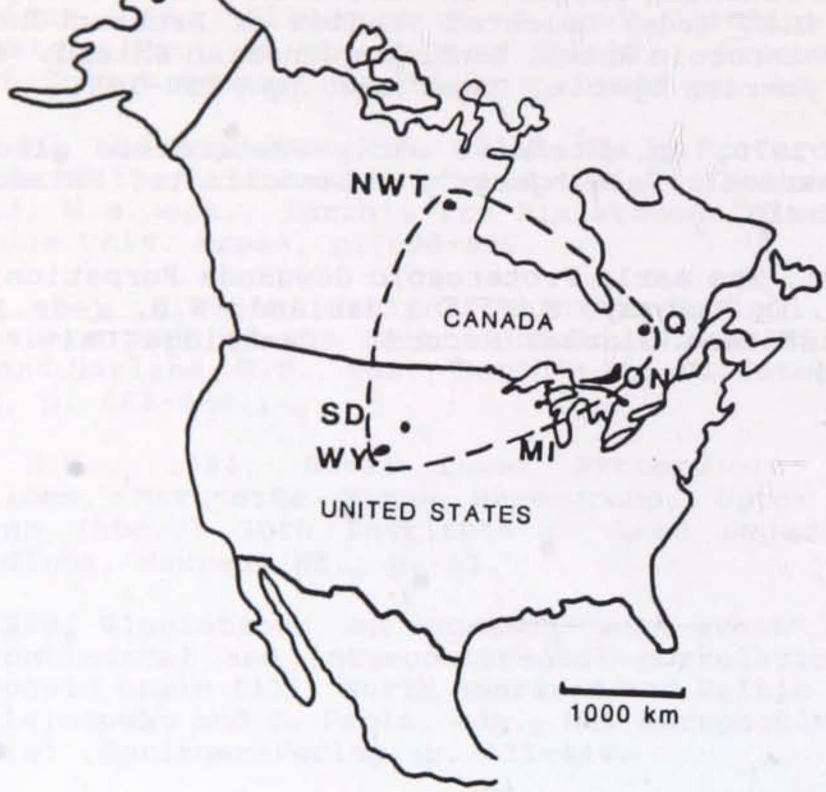


**Figure 1.** Map showing distribution of localities with Early Proterozoic glaciogenic rocks in Ontario (ON), Michigan (MI), Wyoming (WY), Northwest Territories (NWT), and Quebec (Q). also shown is the diamictite locality of the Black Hills, South Dakota (SD). Area enclosed by dashed line was suggested as the area of Early Proterozoic glaciation by Young (1970).


**Figure 2.** Generalized map of the Baltic Shield. Precambrian rocks are shaded; Phanerozoic rocks are white. Numbered black dots represent localities where rocks interpreted to be glaciogenic (this paper) are present. Open dots in USSR are other diamictite localities described by Negrutsa and Negrutsa (1981 a,b). Localities are as follows: 1, Koli-Kaltimo; 2, West side Kontiolahti dome; 3, Sarkilampi; 4, Vayrylankyla; 5, Kurkikyla; 6, Naama; 7, Elmusjoki; 8, Kumsa; 9, Padun; 10, Kalevolampi; 11, Syvatnovolok; 12, Kreznaya Rechka; 13, River Luzhma.

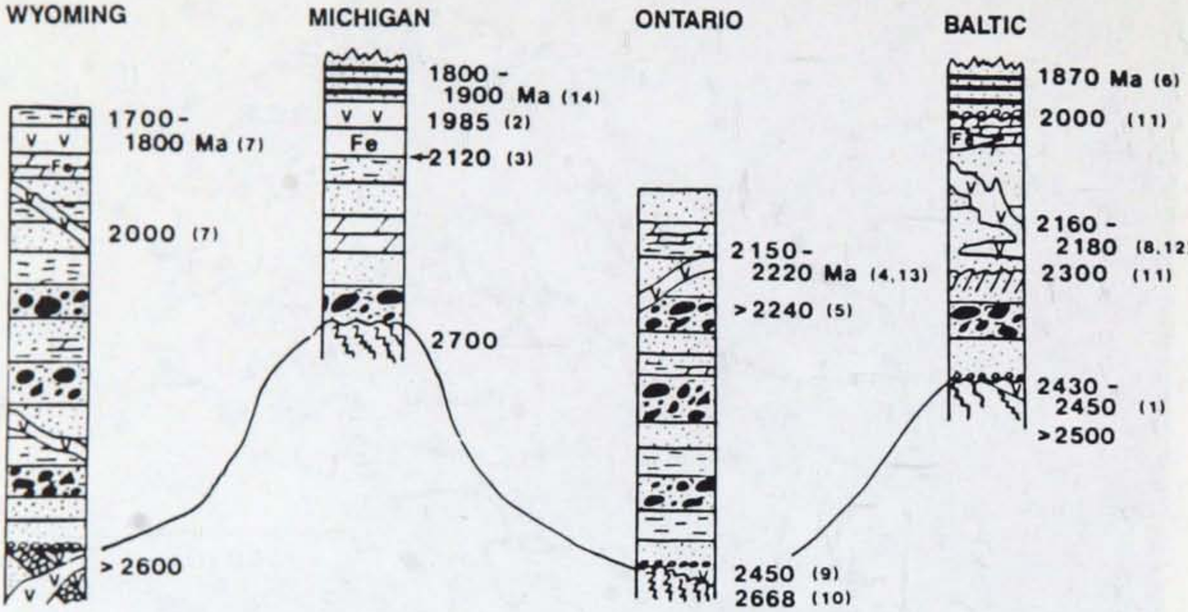


**Figure 3.** Generalized correlation chart for Early Proterozoic columns of North America and the Baltic Shield (Finland and Karelia, USSR). The black conglomeratic pattern represents glaciogenic units, Fe represents iron-formations, and dashes with solid black lines represent turbidite sequences. Other patterns are standard geologic patterns for volcanic rocks, conglomerate, sandstone, shale, and dolomite. Straight lines are conformable contacts and wavy lines are unconformities. Regolith in Baltic column is shown by diagonal lines. Not to scale. Numbers are radiometric dates based on the following references: (1) Alapieti 1982; (2) Banks and Van Schmus 1971; (3) Beck and Murthy 1982; (4) Corfu and Andrews 1986; (5) Fairbairn et al. 1969; (6) Huhma 1986; (7) Karlstrom et al. 1983; (8) Krats et al. 1976, in Salop 1983; (9) Krogh et al. 1984; (10) Krogh and Turek 1982; (11) Merilainen 1980; (12) Sakko 1971; (13) Van Schmus 1965; (14) Van Schmus 1980.


**Figure 4.** Generalized map showing relationships of areas with Early Proterozoic glaciogenic deposits (black) on Canadian and Baltic Shields. Note that the Baltic Shield has been rotated 90° and moved closer to North America. Base after Piper, 1983.

# PRECAMBRIAN STRATIGRAPHY OF KARELIAN DEPOSITS

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The Karelian region located in the SE Baltic Shield is one of the regions favourable for the study of the Precambrian. It covers a time span of no less than 3000 Ma in Precambrian geologic history and is notable for the deep-seated zones of the Earth's crust exposed by erosion. The region is dominated by rock complexes ranging in age from 1600 Ma to more than 3000 Ma and is overlain by relatively thin Quaternary strata. Riphean, Vendian and Paleozoic rocks have only been reported from the south and southeast (Fig. 1).

Three areas are recognized in Karelia in terms of both geological structure and evolution in Precambrian time:

1. The White Sea area is adjacent to the western White Sea coast and occupies part of the Kola Peninsula in the north.

2. The Ladoga area, which lies SW in the Lake Ladoga region, is bounded by Paleozoic deposits to the south and occupies part of Finland in the west-northwest.

3. The Karelian area occupies the rest of the region and is bounded by Paleozoic deposits on the southeast edge of the Baltic Shield.

According to both geological and geophysical data, the above areas form part of the large White Sea, Ladoga and Karelian geoblocks distinguished in the present structure of the Earth's crust in the European Soviet Union (1). They differ in crustal thickness, supracrustal sequences and both endogenic and exogenic activity patterns. Also, they are characterized by seismic, density and magnetic heterogeneities of the lithosphere.

The Karelian part of the White Sea geoblock is a folded sequence known as the Belomorides in the literature. It is a relatively well-preserved portion of an early Archean (Saamian) fold belt.

Structurally, the Karelian geoblock is most complicated on a regional scale. It is built up of geological complexes that formed in structural stages and vary in age from early Archean (Saamian) to Late Proterozoic (Vendian) and Paleozoic.

The Ladoga geoblock forms part of the Svecofennian fold belt which developed in late Early Proterozoic both here and to the west in Finland, Sweden, and Norway.

Elaboration of the principles of stratigraphic subdivision of the Precambrian in the region is based on the use of methods for comprehensive geological and historical investigation, techniques and principles of stratigraphic subdivision used for Phanerozoic terrains, and absolute age determination.

In the regional stratigraphic scale (Fig. 2), the following stratigraphic units are recognized:

1. Complex. A first-rank unit composed of heterogeneous rocks. Its boundaries are structural, related to diastrophism (folding, metamorphism and granitization) epochs and subsequent peneplanation.

2. Subcomplex. A second-rank unit which consists of heterogeneous strata formed at a certain evolutionary stage of the complex. Its boundary is indicated by local nonconformities, conglomerates and crusts of chemical weathering.

3. Superhorizon. A third-rank unit composed of strata which reflect a large transgressive-regressive cycle in sedimentation and volcanism. Its geological boundaries are indicated by discordances and gaps emphasized by crusts of weathering and pronounced scour.

Other units commonly used are:

1. Series. The largest local stratigraphic unit comparable in size to a complex or superhorizon on the regional scale. It can be divided into suites and has its own geographic name.

2. Suite. A local stratigraphic unit recognized with due regard for facies and lithological characteristics. It is bounded by either a structural-facies zone or another part of a geological region. It comprises the deposits formed in the region under certain physico-geographical conditions and occupies a definite position in the above region. It is divided into lower, middle and upper subsuites and has its own geographical name. It is closest to the concept of "formation" of international usage.

The following complexes are distinguished in the Precambrian of Karelia: Saamian, Lopian, Karelian, Riphean and Vendian. The Saamian, Lopian, Karelian and Riphean complexes represent major (600-900 m.y. long) stages in the geological evolution of the Earth's crust. They roughly coincide in both volume and boundaries with global Earth's crustal deformation epochs that are remarkable for maximum heat generation during the periods of over 3.5, 2.8-2.6 and 2.0-1.7 Ga (2).

According to the general stratigraphic scale adopted in the Soviet Union, Saamian and Lopian complexes distinguished on the regional scale correspond to the Archean (3), and the Karelian, Riphean and Vendian complexes correspond to the Proterozoic. Everywhere the Archean-Proterozoic boundary is geologically emphasized by a sharp angular unconformity which can be traced by mapping. The erosion at the boundary was deep enough to have reached granites. This is indicated by the Early Proterozoic (Sumian) crust of physical weathering (breccia), basal conglomerates and sandstones that developed on both the granites and the Lopian complex.

In the White Sea block, the early Archean (Saamian) complex is represented by the gneisses and amphibolites of the Belomorian series. Their composition and structure remain uniform for hundreds of kilometres along strike. There are no reliable textural and structural relics indicative of the protoliths. In the Karelian geoblock, Saamian rocks occur as relics in the granite-gneiss areas of Lopian anticlinal structures, where they are present as amphibolites, bipyroxene crystalline schists, gneisses and gneissose granites. The volcanogenic (basaltic and komatiitic) nature of the Volotsk rock sequence (4), regarded as Saamian, has only been established at one locality in the SE part of the block. The Volotsk rocks have been dated at 3391 $\pm$ 76 Ma by the Sm-Nd method (D.Z. Zhuravlev and N.S. Pukhtel, IGEM, USSR Academy of Science), whereas the cross-cutting tonalites have been dated at 3540 $\pm$ 60 - 3500 $\pm$ 90 Ma by the U-Pb method (5).

The late Archean (Lopian) complex contains conglomerates, quartzites, magnetite quartzites, carbonaceous shales and volcanics that differ in composition. The complex differs sharply in this respect from adjacent amphibolite-gneiss deposits. Primary nature, depositional setting and depositional sequence are ascertained for all the rocks. The Saamian-Lopian boundary has an age of no less than 3200 Ma (6) and probably no more than 3500 Ma.

The Early Proterozoic (Karelian) complex falls into three subcomplexes. They are separated from each other by both angular unconformities and structural boundaries and differ in intrinsic sets of sedimentary and volcanogenic strata. The lower subcomplex is dominated by Sumian volcanogenic rocks and Sariolian conglomerates. The middle subcomplex, with a crust of chemical weathering, consists mainly of (a) alternating Jatulian volcanogenic and sedimentary strata represented largely by quartz conglomerates, quartzites and dolomitic rocks with stromatolite-oncolite bioherms; and (b) Ludicovian deposits built up by volcano-sedimentary, carbonate and chemogenic siliceous strata in which different types of carbonaceous (shungitic) rocks are widespread. The upper subcomplex is represented by Kalevian sandy-argillic flyschoid sediments in the Lake Ladoga region (Ladoga series) and by Vepsian sandy-quartz rocks and red beds in the Karelian geoblock.

The Late Proterozoic (Riphean) complex consists of the Middle Riphean red and mottled sandstone, gravelstone, conglomerate and sedimentary-volcanogenic rocks (basalts, dolerite basalts and tuffs) of the Salmi suite, which rest unconformably on the crust of weathering developed on rapakivi granites and are overlain with angular unconformity by Vendian deposits.

Vendian rocks, which constitute the upper part of the Precambrian sequence in Karelia, were earlier referred to as Lower Cambrian. They rim the southern Baltic Shield, fill marginal grabens and are overlain by the Russian Platform cover sediments. They are conglomerates, gravelstones, sandstones, siltstones, argillites and clays of the Valdai series, which rest with sharp angular unconformity on a thick crust of weathering of all the Precambrian rocks of the region.

It should be noted in conclusion that the present areal distribution of the above complexes as well as their understanding and the lateral tracing of their boundaries are not uniform. Absolute age determination is widely used for their dating and correlation although a comprehensive approach to the study of Precambrian deposits will, no doubt, require the further use of both geological and paleontological methods.

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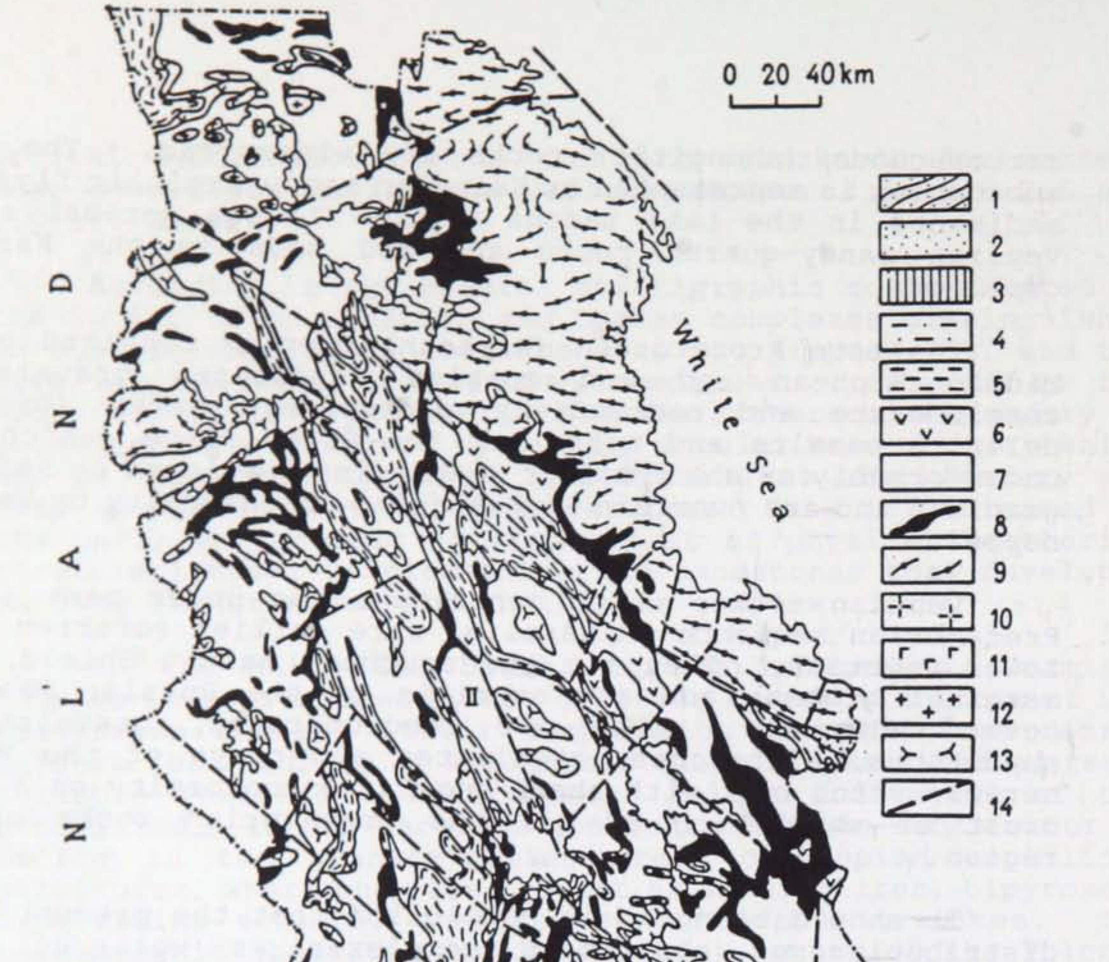


Fig. 1. Geological structure of Karelia.

Paleozoic: 1 - Carboniferous and Devonian systems. Late Proterozoic: 2 - Vendian, 3 - Middle Riphean. Early Proterozoic - Karelian: 4 - upper subcomplex, 5 - middle subcomplex, 6 - lower subcomplex. Archean: 7 - individed Archean (partly Early Proterozoic), predominantly granitoid and migmatite deposits. 8 - Lopian, 9 - Saamian. Proterozoic intrusions of: 10 - rapakivi granites, 11 - basic and ultrabasic rocks, 12 - granites and granodiorites. 13 - Archean granite and granodiorite intrusions. 14 - faults. I - White Sea geoblock, II - Karelian geoblock, III - Ladoga geoblock.

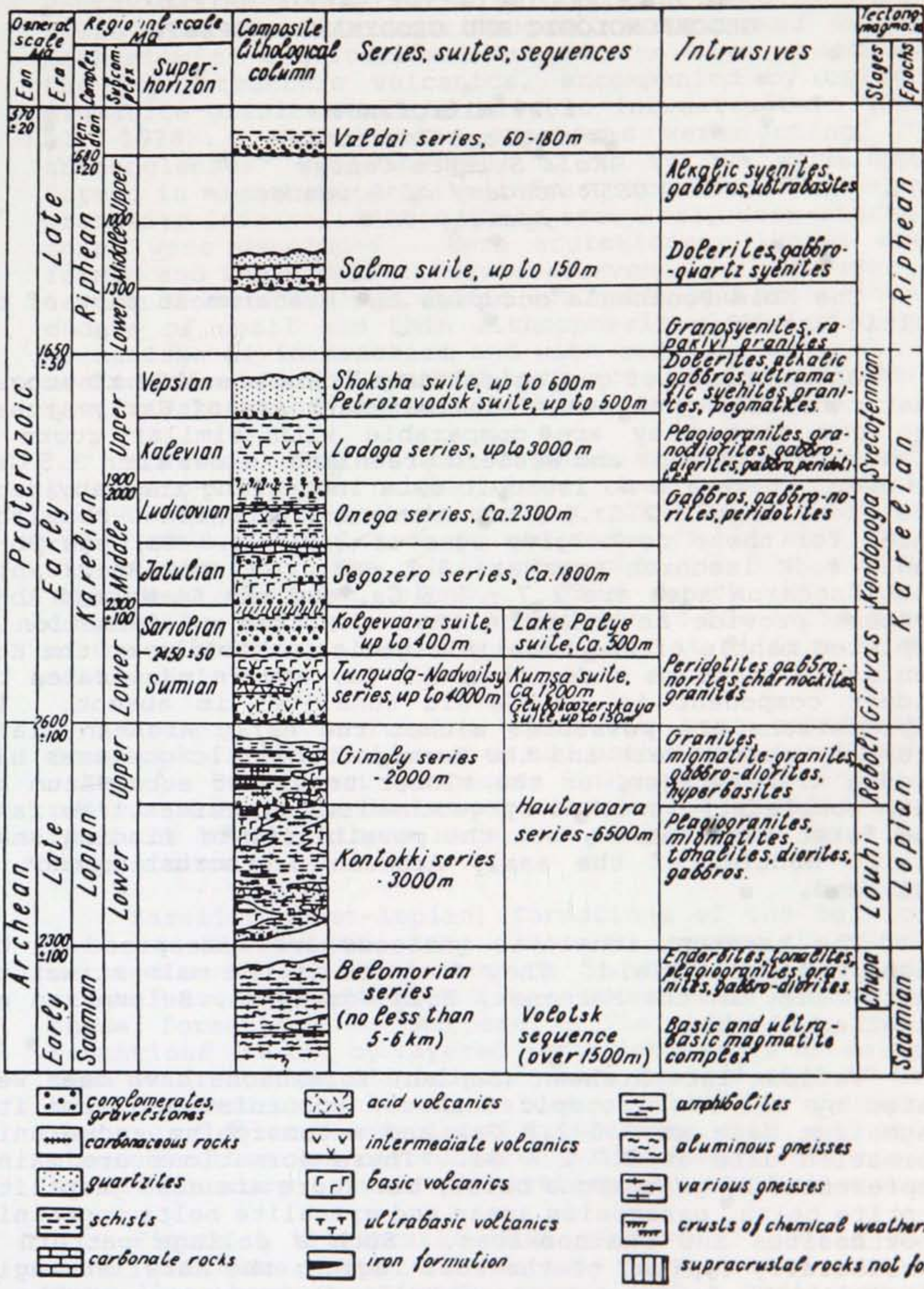


Fig. 2. Precambrian stratigraphy of Soviet Karelia.

# ARCHEAN AND EARLY PROTEROZOIC GEOLOGY OF THE KOLA REGION OF THE BALTIC SHIELD: GEOCHRONOLOGIC AND GEODYNAMIC ASPECTS

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The Kola Peninsula occupies the easternmost part of the Baltic Shield.

On the basis of geological data it is possible to suggest that tonalitic gneisses of the basement are of Early Archean age and that they are comparable with similar rocks of Minnesota, Labrador and Western Greenland (older than 3.5 Ga). However, there are no isotopic data indicating that any rocks are older than 3.2 Ga. Concordant and isochron U-Pb zircon dates for these rocks give ages of 2.9 - 2.8 Ga, the Pb-Pb whole rock isochron ages are 3.2 - 3.0 Ga, the Rb-Sr whole rock isochron ages are 2.7 - 2.6 Ga, and the Sm-Nd and Rb-Sr systems provide no evidence that the time of formation of depleted mantle of the tonalitic gneiss protolith on the Kola Peninsula might be older than 3.2 Ga, and in many cases the oldest component in 2.9 Ga-old tonalites is absent. Two explanations are possible: either the early Archean sialic crust is absent here and the Saamian tonalitic gneisses have middle Archean ages, or the oldest tonalitic substratum has been completely reworked by geochemical processes. We favor the first explanation, but the possibility of finding small sialic nodules of the early Archean protocrust cannot be excluded.

The basement tonalitic gneisses are widespread in the eastern Baltic Shield. They dominate in the main structural units, i.e. in the Murmansk, Kola-Norwegian, Belomorian and Karelian blocks.

Various late Archean (Lopian) formations have been well dated by all the isotopic methods; volcanism and tonalitic magmatism date at 3.0-2.8 Ga, and metamorphism and granite formation date at 2.7-2.6 Ga. These formations are mainly represented by greenstone belts, but there are also jaspilite-leptite belts, paragneiss areas and granulite belts containing anorthosites and charnockites. Such a collage pattern is particularly typical of the Kola region; the Karelian region to the south is more homogeneous. It seems reasonable to consider the Kola region to be "an Archean high-grade metamorphic domain" (Windley, 1977), or "Archean granulite-greenstone area" (according to our terminology).

As to the origin of typical greenstone belts (e.g., the Hautavaara Complex in Karelia), we tend to conclude that these belts are of ensialic riftogenic origin and we assume a mechanism of gravitational subsidence (i.e., sagduction) of the suprastructure volcanics, accompanied by uplifting of migmatite-granites of the plastic infrastructure (Gorman et al, 1978). While these processes were acting, "crustal asthenolenses" with anatexis of up to 30% were apparently formed in migmatite-granite infrastructures. In charnockite-granulite infrastructures (Kola examples), deep-seated thrust zones were developed. Here accretionary lenses were also formed and there were obductional events in the lower crustal layers of anorthosites, mafic coronites and granulites. Such models of small and thin lithospheric plates with peculiar mechanisms of interaction and with crustal asthenolenses and heated lower crust layers seem to account for the complicated mosaic pattern of the Kola Archean formations.

However, some of the Archean greenstone belts (e.g., Kolmozero-Voronja) might have had an ensimatic and intercontinental tectonic setting. This assumption cannot be ruled out. In any case, it can serve as the most reasonable explanation for the alien character of the northernmost (Murmansk) Archean block. This block is similar in composition and evolutionary features to the Archean terrains of Greenland (e.g., Amitsok area), but it differs from blocks of the Kola-type.

Archean crust-forming processes have constructed the framework for the Kola craton and especially for the Karelian craton. Subsequent processes only resulted in a partial restructuring. The latter was accompanied by mafic intrusions from the mantle; these made the sialic crust more mafic in composition.

Karelian (post-Lopian) formations of the Baltic Shield can be subdivided into several groups. The lower metasediments and metavolcanics of the Lapponian and Sumian stages have ages of 2.6-2.5 Ga. It would be better to regard these formations as Archean. The point is that these formations are cut by layered mafic intrusions whose ages have been determined by all isotopic methods as $2.45 \pm 0.05$ Ga. They have constant negative $E_{Nd}$ values. This is a characteristic feature of a large undepleted mantle reservoir, and shows that the convection was restructured 2.45 Ga ago, or approximately at the boundary between the Archean and the Proterozoic.

Figure 1 illustrates that sources of mantle magmatism were the same, and the magmas that formed the layered intrusions have negative $E_{Nd}$ values. This indicates that the magmas belong to the undepleted mantle. This fact distinguishes them from the Archean tholeiites and komatiites. The Baltic Shield is rather interesting in terms of homogeneous properties of all the layered intrusions that occupy an extensive area. It is difficult to explain this fact by the effect of crustal contamination. This distinguishes layered intrusions of the Baltic Shield from the Sudbury layered intrusion which is contaminated.

Early Proterozoic volcanites and Jatulian sediments of various compositions are 2.4-2.0 Ga old. On the Kola Peninsula they compose the Pechenga-Varzuga intracratonic riftogenic greenstone system. It is possible to speak about several stages of activation of the mantle diapirs, rifting processes, magmatism and sedimentation, and related intracratonic collisional compression and metamorphic events.

To understand the nature of collisional geodynamics of the region, it is very important to study the inter-relations of fold systems of the Finnish Svekofennides, the Karelian and Kola Karelides, and the Belomorides. Their interpretation on the basis of various geophysical data is presented in the Third European Geotransect.

The Karelian craton with its oldest geological formations (more than 3.1 Ga old) and well-preserved Archean and Proterozoic greenstone belts has the simplest deep structure and the thinnest crust.

A "crust-mantle mixture" has been established beneath the Svekofennian ensimatic folded area of the final stages of the Early Proterozoic (2.0 - 1.7 Ga). This mixture has peculiar velocity and density properties: 7.0 - 7.4 km/s and 3.0 - 3.2 g/cm$^3$. This lower crustal layer has a maximum thickness in the Outokumpu region of Finland. Finnish geologists G. Gaal and A. Tuomi have determined the presence of an allochronous obducted ophiolitic complex. Its age is 1.96 Ga and the crust in this area is anomalously thick, as much as 60 km. A general increase in the crustal thickness can be explained by subduction of the Svekofennian oceanic crust under the active margin of the Karelian continent.

The hinge zone between the Karelian craton and the Belomorian block has interesting characteristic features. In the modern section it is a typical shear-structure, with blastomylonites formed from intrusive charnockites. In addition, ultramafic rocks and mafic and alkaline intrusions are present. This suture is characterized by crust as thick as 50 km, by the presence of the "crust-mantle mixture" at the base, and by increased electrical conductivity and modern seismic activity.

The Kola domain at depth is interpreted as having a very complicated construction with "cogged" structure of the Earth's crust. There are a lot of rocks having properties of granulites, mafics and ultramafics. They compose inclined plate alternation with various but always increased densities. The deep tectonic melange of mantle, lower crustal and supracrustal rocks is mainly characteristic of the Lapland Granulite Belt. It is interpreted as an intracratonic abyssal obducted structure which is associated in its evolution with the Belomorides.

The Belomorian granulite-gneiss area occupies an axial position in the oldest part of the Baltic Shield. From deep seismic records it has been found that this area is notable for its numerous subhorizontal reflecting horizons that are concordant with deep-seated thrusts of Lapland. This seismic heterogeneity is interpreted as tectonic bedding related to deep thrust plates and shear zones. This type of geodynamics is supported by geological indicators such as high-pressure metamorphic rocks, basic coronites and anorthosites, and muscovite pegmatites.

High-pressure metamorphism was repeated in the Belomorian block for more than a billion years, in the time span from 2.9 to 1.8 Ga. However, this high-pressure metamorphism can be related to a collisional event only during the closure of the Svekofennian paleo-ocean 1.9 to 1.8 Ga ago. Several geological data sets necessitate a different mechanism to explain the older high-pressure metamorphic events. We propose such a mechanism and call it "a shut-down mechanism".

Approximate estimates of gravitational paleofields for the three major blocks of the mobile craton at 2.4 Ga ago, the beginning of the Jatulian, provide varying data: Karelian block -10 to -15 mGal, Belomorian block +20 to +30 mGal, and the Kola block +10 to +15 mGal. This gravitational difference should have been reflected in the gradient subsidence and in gradual shutting down of the heavier Belomorian mass by adjacent masses that were lighter. As the T-P gradients were increasing, the following phenomena could have taken place: active metamorphic processes along shear zones, mass-exchange within the crust, and mass-exchange between the crust and the lithospheric mantle.

Thus, the eastern part of the Baltic Shield during its early Precambrian history was not a single entity, but a collage structure composed of small landmasses or terrains. They have various geological aspects, including ore deposits, and different geological evolutions. This is illustrated in the tables for the three large structural blocks, the Kola, Belomorian and Karelian.

In conclusion, it is interesting to note that 50 years ago academician Alexander Fersman singled out the Kola Alkaline Province with its unique deposits of alkaline rocks (Khibiny, Kovdor, Lovozero) as Paleozoic in age. It has now been established that this province originated as such 2.4 Ga ago. This fact also distinguishes the Kola region from the Karelian and the Belomorian blocks.

# LATE ARCHEAN GREENSTONE BELTS IN KARELIA

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The classic Karelian granite-greenstone region of the eastern Baltic Shield occupies an area of about 125,000 km$^2$. Structurally, it consists of 1) blocks of rheomorphosed basement ($Ar_?$) represented by tonalite, granodiorite, trondhjemite, gneissose granite, and young granite; 2) late Archean (Lopian) greenstone structures; and 3) Proterozoic cover.

Seven greenstone belts -- the Sumozero-Kenozero, South Vygozerc, Parandovo-Tiksheozoro, Vedlozero-Segozero, Gimoly-Kostomuksha, Jalonvaara-Ilomantsi-Tulos and Kuhmo-Suomussalmi belts -- are recognized in the Karelian granite-greenstone region, which includes both Karelia and eastern Finland. These late Archean greenstone belts (ca 3.3 - 2.7 $\pm$ 0.1 Ga) are represented in the present section by numerous local relict structures that vary considerably in size. The sedimentary rock sequences within the belts can be traced as far as hundreds of kilometers. The widths of the structures vary from 10 to 20 km (Fig. 1). On geophysical maps they appear as positive gravity and magnetic anomalies.

The greenstone belts are controlled by a conjugate system of orthogonal and diagonal deep faults (mobile-permobile zones) which cut the sialic crust into blocks of different orders. On gravity maps, the blocks are marked by both morphology and the intensity of regional gravity anomalies. The mobile-permobile zones which separate the blocks developed simultaneously with the block structures and are indicated by intense horizontal step-like gravity gradients. The thickest volcanic strata, relics of eruption centers and ultrabasic to basic intrusion belts, are confined to the mobile-permobile zones.

Greenstone belts fall into two types depending on the rank of controlling faults. The first type, associated with intergeoblock and intrablock deep-fault zones, is characterized by the most intense and compositionally variable magmatic rocks, subordinate sedimentary rocks, the thickest volcanic strata, and numerous relics of eruption centers. This type comprises all belts except the Gimoly-Kostomuksha and the Jalonvaara-Ilomantsi-Tulos belts. These two belts belong to the second type. They are controlled by intrablock

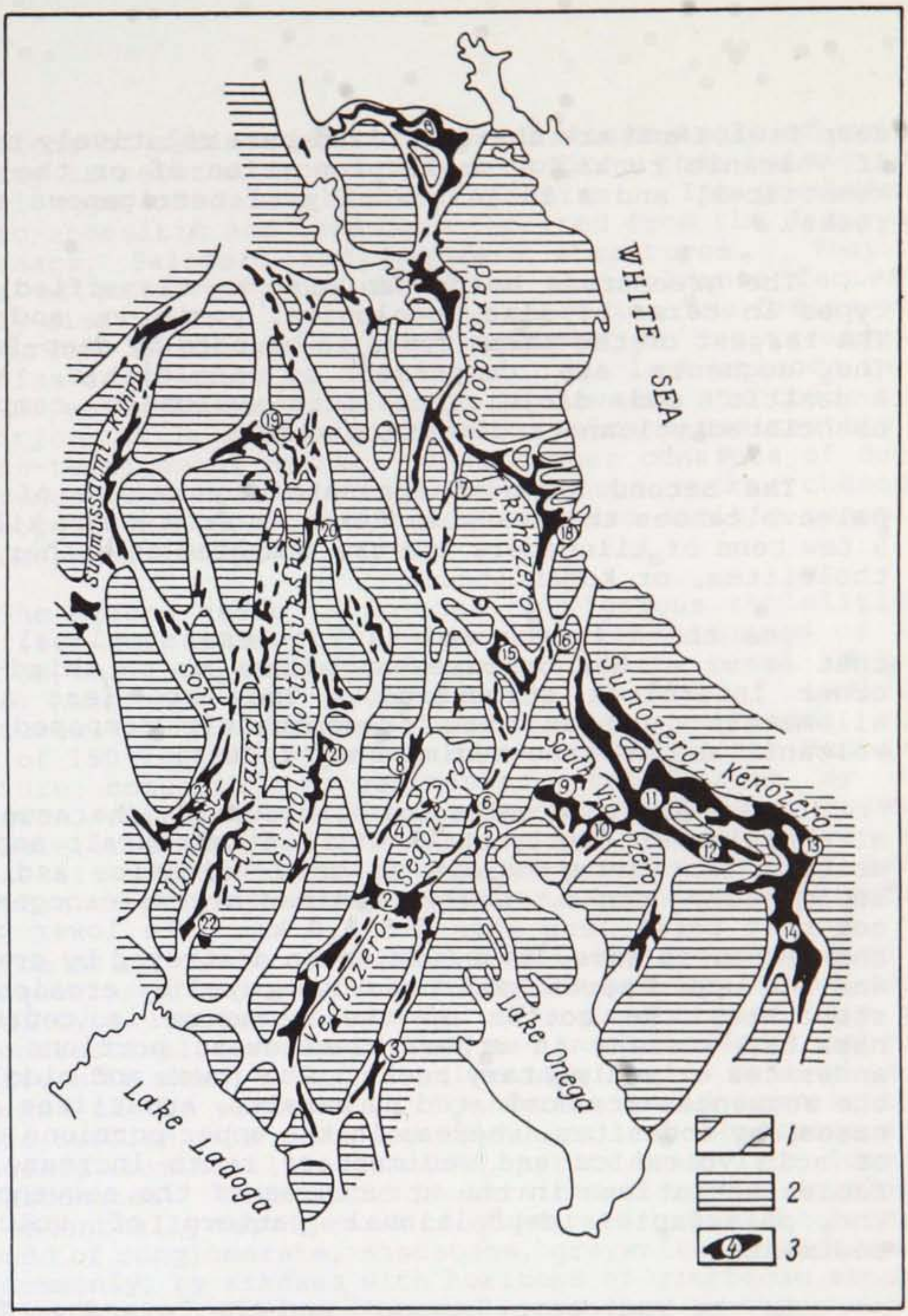


Fig. 1. Reconstruction of Karelian greenstone belts. 1 - greenstone belts; 2 - granitoid blocks of rheomorphosed basement; 3 - greenstone structures (numbers on scheme): 1 - Hautavaara, 2 - Koikary - Korbozero, 3 - Manga, 4 - Sovdozero, 5 - Palaselga, 6 - Saiozero, 7 - Bergaul, 8 - Luzhmozero, 9 - Shilos, 10 - Rybozero, 11 - Lake Kamennoye, 12 - Senegozero, 13 - Kozhozero, 14 - Toksha, 15 - Parandovo, 16 - East Idel, 17 - Tunguda, 18 - Pebozero, 19 - Kostomuksha, 20 - Kedrozero - Bolshezero, 21 - Gimoly, 22 - Jalonvaara, 23 - Ilomantsi.

deep faults and are characterized by a relatively small volume of volcanic rocks, a small proportion of or the absence of komatiites, and a large quantity of terrigenous sedimentary rocks.

The greenstone belts can also be classified into three types in terms of size, geological position, and structure. The largest of the three types is bounded by deep-fault zones. The segments are comprised of tholeiitic, komatiitic, andesitic and dacitic-rhyolitic volcanic complexes and associated volcano-sedimentary strata.

The second type is dominated by relics of individual paleovolcanoes that vary in diameter from a few kilometers to a few tens of kilometers and are composed of either andesites, tholeiites, or komatiites.

The third type combines rootless synclinal structures that occur either on basement blocks or on third-to-fourth-order interblock structures. They are less than a few kilometers long and are predominantly composed of distal volcanic and volcano-sedimentary facies.

The greenstone units are structurally characterized by an alternation of fairly thick (to 2.5 km) basalt and komatiite members with intermediate to acid volcanics and associated sedimentary rocks. The sedimentary-volcanogenic strata commonly total less than 5.5-6.0 km. The lower portions of the sequences have, as a rule, been destroyed by granitization and the upper parts have been destroyed by erosion. In some structures the bottom of the sequence is represented by basalts, whereas in others the lowest portions consist of andesites or sedimentary rocks. The lower and middle parts of the sequences are dominated by basalts, komatiites and in some cases, by andesites, whereas in the upper portions the amounts of acid volcanics and sedimentary rocks increase. Lateral facies variations in the structures of the sequences reflect the paleofacies depositional pattern of volcanics and sediments.

In the Vedlozero-Segozero and the Parandovo-Tiksheozero greenstone belts of Central and East Karelia, intermediate, basic and ultrabasic volcanics are widespread, whereas sedimentary rocks are less common. In the Gimoly-Kostomuksha belt (with the exception of the Kostomuksha structure of West Karelia), the stratigraphic sequences are not as thick, the proportions of terrigenous sedimentary rocks and acid volcanics are greater, the volume of basalts is markedly less, and no komatiites are observed.

As a whole, the greenstone belt sequences show a cyclicity and contain numerous unconformities with the latter indicated by either conglomerates or weathering crusts (i.e., (paleosols). Five cycles of volcanic activity are separated by periods of sedimentation. The earliest cycle of volcanic activity is related to volcanic eruptions that belong to the central type of calc-alkalic volcanism. Its products are dacitic-andesitic and have been reported from the Hautavaara, Jalonvaara, Saiozero and Pebozero structures. They form relics of paleovolcanoes in which the volcanic facies are zonally distributed. Diatreme and near-diatreme facies occupy an area of 15-25 $km^2$ and the products of dacite-andesitic volcanism cover areas of 75-190 $km^2$. The last episode in this cycle of volcanic activity in the Hautavaara structure is the deposition of a 100-150 m-thick volcano-sedimentary and volcano-terrigenous member. This member consists of dacitic sandstones, sulphide-bearing carbonaceous rocks, chemogenic siliceous rocks, quartzose sandstones and dacitic monomictic conglomerates and gravelstones.

The second cycle is related to vigorous tholeiitic and komatiitic volcanism. It is essentially composed of lava, contains tuff and sedimentary rocks (chert, black shale and pyrite) horizons and accounts for 40-80% of individual structures. Tholeiites, relics of shield volcanoes, occupy areas of 150-200 $km^2$. The top of the cycle in the Hautavaara structure consists of sedimentary rocks such as iron-formation, sulphide ores, carbonaceous shale, and graywacke. The tholeiitic-komatiitic volcanism was separated from later calc-alkalic volcanism by a time of primary folding and the formation of polymictic conglomerate, graywacke and arkoses.

The third cycle of volcanic activity is associated with new sequences of calc-alkalic volcanism and the development of central-type insular volcanoes. The products of this stage are collectively referred to as andesite-dacite-rhyolite formations in which volcano-sedimentary facies make up as much as 20-25% of the sequences. The formation is represented by dacitic tuffite (i.e., eroded volcanics), carbonaceous shale, chemogenic aluminosiliceous rocks, pyrite ores, iron-formation and carbonates. This cycle is terminated by the development of an essentially terrigenous conglomerate-graywacke formation composed of conglomerate, sandstone, graywacke, siltstone and, less commonly, by arkoses with horizons of quartzose sandstone and black shale. The terrigenous rocks vary in thickness from tens of meters to a few hundred meters in different structures.

The fourth komatiite-tholeiite cycle of volcanism is petrochemically similar to the second cycle. It is best developed in the Hautavaara, Kostomuksha and Hizovaara structures.

Data on the fifth cycle of volcanic activity are scarce. In the Hautavaara structure it is represented by the 150-200 meter-thick Usmitsanjarvi suite which consists of tuff, tuffite, graywacke, aluminosiliceous rocks and carbonaceous shale.

Thick flysch-type sedimentary rocks (turbidites), which are separated from basalts at the base by conglomerate and a weathering crust were probably formed at the same time in western Karelia. The bulk of iron-formation in Karelia is associated with the flysch-type sedimentary rocks.

A few major stages of metamorphic and metasomatic transformations are recognized in the greenstone structures:

1. Autometasomatism of volcanic complexes during both volcanism and post-volcanism stages.

2. Subsidence metamorphism related to an early folding phase.

3. Zonal regional metamorphism associated with a major folding phase and granitization. It is concomitant with various metasomatic alterations.

4. Superimposed metasomatic alteration associated with Karelian and Svecofennian tectonogenesis.

# EARLY PRECAMBRIAN HIGH-MAGNESIAN MAGMATISM IN THE BALTIC SHIELD

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High-magnesian magmatism is a major source of information on the Earth's deep shells, and is commonly used when elaborating upon geodynamic models for the evolution of the Earth and its large regions. Some generalizations for the ultramafics of the Baltic Shield (1, 2) have shown high-magnesian magmatic rocks (over 9% MgO) to be in 10 Precambrian stratons which rank as superhorizons in the regional stratigraphic scale. One occurrence is in the Saamian ($AR_1$) complex, three are in the Lopian ($AR_2$), five are in the Karelian ($PR_1$, 1-3) and one is in the Riphean ($PR_2$).

The axial lines of the corresponding zones are indicated on the distribution schemes for high-magnesian magmatites and associated strata and the evolutionary patterns of mantle magmatism are generalized (Fig. 1). Most of the patterns are characterized by the following geometric types: polygonal, orthogonal and linear. No type has so far been established for $AR_1$ due to the lack of evidence. Pukhtel et al. (unpublished results) have recently reported a Sm - Nd age of 3.39 $\pm$ 0.076 Ga for the Saamian komatiites which occur in basalts metamorphosed under amphibolite-facies conditions in the Vodlozero block, east of Lake Onega. A Saamian age is assumed, but not confirmed by geochronological methods, and a similar age is assumed for the ultramafics that occur together with the amphibolites of the Hotolambina suite of the western White Sea region.

In $AR_2$ (3.1-2.9 Ga), high-magnesian magmatism of predominantly komatiite and partly tholeiite series is related to the riftoid developmental stage of 5 to 7 km-thick Lopian greenstone belts in which high-magnesian volcanics account for 3.12% (2). Intrusive comagmatic rocks are usually represented either by ultramafic dikes in the basement of the belts (e.g., the Vinelskaya dike in the Vodlozero block) or by sills and stocks (e.g., the Kamennoozerskaya structure) within the belts. Some contain Cu-Ni ore deposits (e.g., Allarechka, Vozhma). The geometric pattern of riftoid zones, in which high-magnesian comagmatic rocks are present, is the polygonal type (Fig. 1 a). Blocks with weak or no mantle magmatism are regarded as relics of the Lopian lithospheric microplate. The polygonal type of mafic-ultramafics ($AR_2$) is also well-defined in the mid-Dnieper and Kursk granite-greenstone areas of the

East European platform (3).

In $PR_1$ (2.5-2.4 Ga), high-magnesian magmatism was apparent at the protoaulacogen evolutionary stage of the Fenno-Sarmat protoplatform. The volcanics of the komatiite series have been mainly reported from the Pechenga-Varzuga, Lapland and East Karelian protoaulacogens, where they account for a few percent of the above structures. The layered peridotite-gabbro-norite terrains, which contain Cr, Ni, Cu, V, and PGE and are widespread in the Kemi-Koilismaa-Kukasozero, Burakovsko-Monastyrskaya, Koitelainen and Pechenga-Moncha-Varzuga, should probably be viewed as intrusive facies of high-magnesian magmas. The above magmatites are thought to belong to either the boninite or komatiite series (4). The evolutionary pattern of high-magnesian magmatic zones approaches an orthogonal type (Fig. 1 b) in which NW- and NE-striking systems are most distinct.

High-magnesian magmatism $PR_1^2$ (2.05-1.97 Ga) manifested itself at the protoplate developmental stage of the protoplatform during the Suisarian riftogenesis epoch in the Vetrany Poyas, Lapland, Pechenga-Varzuga, Onegozersko-Rybinsk and Outokumpu-Kainuu proto-rift structures. High-magnesian volcanics (Fig. 2) of both tholeiite (picrites, picrobasalts and olivine basalts) and komatiite series (komatiites > 24% MgO and komatiite basalts 9-24% MgO) account for 5-50% of the above structures. Intrusive comagmatic rocks are represented by numerous mafic and ultramafic, occasionally nickel-bearing bodies, associated with the tholeiite series (1). The evolution of ultramafics clearly shows a linear type (Fig. 1 c).

In $PR_1^3$ (1.95-1.85 Ga), mafic-ultramafic magmatism was most apparent under intrusive-facies conditions in the Svecofennian province of the Baltic Shield and was more limited in the Kola-Lapland-Karelian province. In the former, mantle magmatites show a polygonal pattern (Fig. 1 d) and in the latter they are confined to the White Sea-Lapland suture (5). The polyphase terrains of the alkaline-ultramafic series with Fe, Ti, P and other occurrences and deposits (Gremyakha-Vyrmes, Yeletozero-Tiksheozero) are restricted to the NNE-striking line, which shifts the above suture and probably indicates a transform fault.

In the Riphean (ca 1.2 Ga), magmatism of the alkaline-ultramafic series is apparent as thin micpicrite veins in Kostomuksha and Vetreny Poyas (Windy Belt) (5).

In the Baltic Shield, high-magnesian magmatites of different series occur in a certain order which shows the following evolutionary trend: the komatiite series is most common in AR, the tholeiite series in $PR_1$ and the alkaline-ultramafic series in $PR_1^3$ and $PR_2$. This reflects corresponding transformations in the mantle (6).

To explain some regularities in the distribution of mantle magmatism products, let us use the convection theory, in which the Raleigh number (Ra) is one of the most important values.

Convection is known to occur if Ra exceeds a critical value. As Ra increases, a convection type is changed by altering the pattern of mantle matter flow from a roller type via bimodal (orthogonal) to a polygonal type (7, 8).

In present geodynamic settings, the intrusion of high-magnesian magmatites takes place in the axial portions of rifts, either above ascending mantle jets or above "hot spots". This could also have occurred in earlier epochs. If the spatial distribution of high-magnesian rocks is regarded as a projection of ascending mantle jets at a certain time, then the early Precambrian evolution of asthenospheric convection can be outlined.

In the Kola-Lapland-Karelian province, the mantle magmatite distribution patterns are simplified from a polygonal type ($AR_2$) to a linear type ($PR_1^2$). This seems to be related to the progressively cooling Baltic segment of the Earth in which the mantle temperature declines, the asthenospheric thickness decreases and both convection type and convection cell sizes are altered (Fig. 3). As indicated by the estimated distance between the axes of ascending jets, the size of the convection cells was 200-300 km for $AR_2$ and 150-200 km for $PR_1^2$. Such a decrease in convection cell size implies that Raleigh number for the Archean asthenosphere was smaller than that for the Proterozoic asthenosphere, the temperature in the latter being lower and other parameters (acceleration due to gravity, density, viscosity, thermal conductance and heat expansion) being equal. This is in good agreement with Raleigh number ratio estimates (1.6-10 and more) made for a transition from polygonal-type convection to a roller type obtained experimentally (8).

The evolutionary trend in mantle convection in the asthenosphere of the Kola-Lapland-Karelian province seems to reflect a characteristic "endogenic cycle" in tectonospheric evolution from AR to $PR_1^2$. A new "endogenic cycle" is observed in the Svecofennian province and the Belomorian megazone in $PR_1^3$. The cycle seems to be associated with additional energy transport from either the lower mantle or the core-mantle boundary to the above segments of the tectonosphere. In early Precambrian time, both energy and matter were transported to the lithosphere through a convecting asthenospheric layer, in which the type of convection was obviously responsible for both the shape and size of microplates or their embryos. The intensity and chemical composition of magmatism at the boundaries of the above microplates are also largely dependent on the dynamics of the microplates (9).

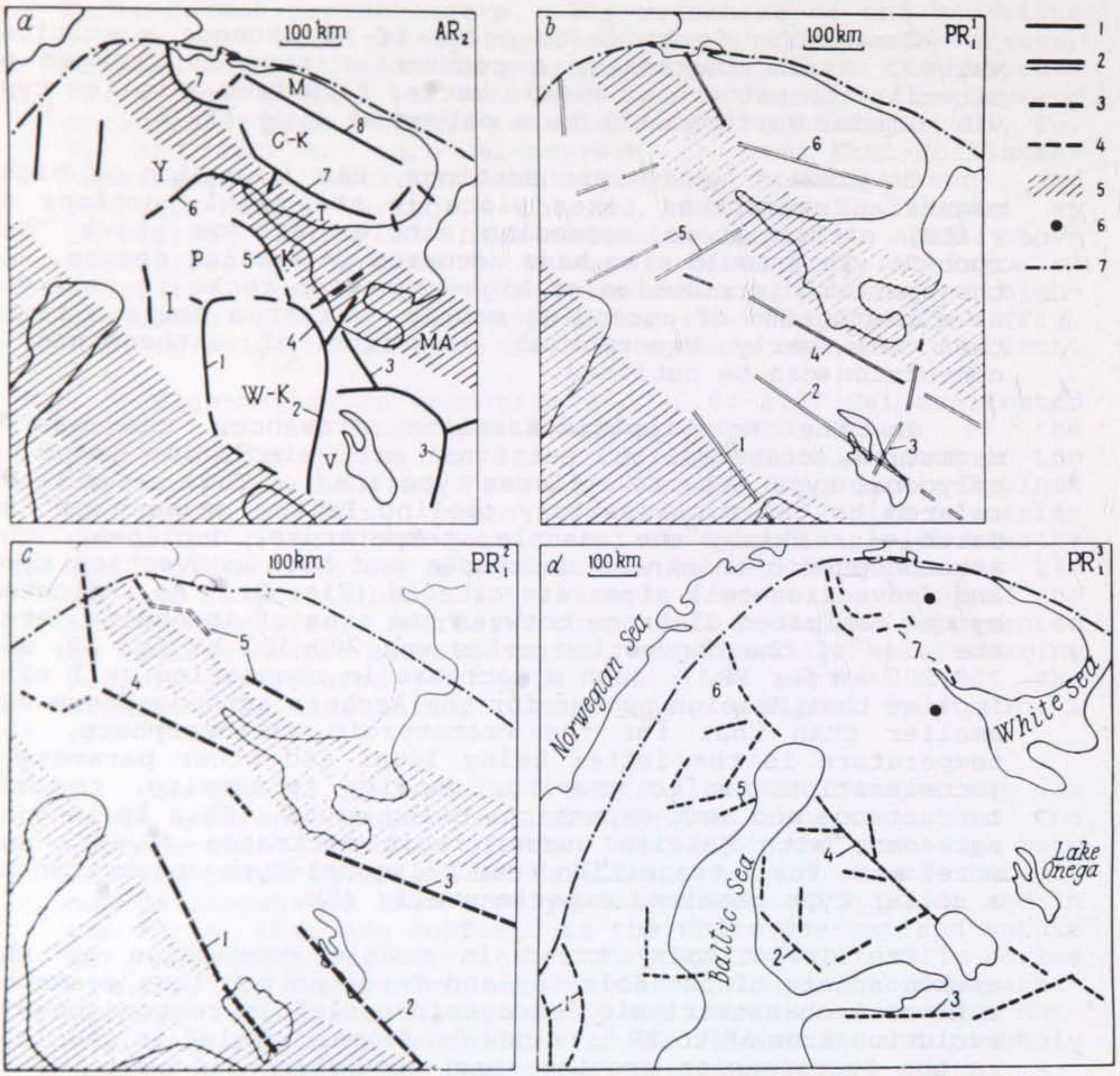
a
100 km
AR2
M
C-K
Y
T
P
K
I
MA
W-K2
V
b
100km
PR11
c
100km
PR12
d
100km
PR13
Norwegian Sea
White Sea
Baltic Sea
Lake Onega

Fig. 1. High-magnesian magmatite zones in $AR_2$(a), $PR_1^1$(b), $PR_1^2$ (c) and $PR_1^3$, Baltic Shield.

$AR_2$: 1-axial zones of high-magnesian magmatism coincident with the following greenstone belts: 1 - Suomussalmi - Kuhmo, 2 - Vedlozero - Segozero, 3 - Sumozero - Kenozero, 4 - Tiksheozero - Pebozero, 5 - Notozero, 6 - Tuntsa - Savukoski, 7 - Tersk - Allarechka, 8 - Kolmozero - Voronya.
Relics of the following microplates: M - Murmansk, CK - Central Kola, Y - Yensk, T - Tersk, P - Pyaozero, K - Kem, Ma - Malenga, I - Iisalmi, WK - West Karelian, V - Vodlozero.

$PR_1^1$. 2 - axial zones of high-magnesian magmatism and protoaulacogens: 1 - Kainuu- Outokumpu, 2 - Central Karelian, 3 - Burakovo - Monastyrsky, 4 - East Karelian, 5 - Kemi - Kukasozero, 6 - Pechenga - Varzuga.

$PR_1^2$. 3 - Axial zones of Ludicovian high-magnesian magmatism and protorifts: 1 - Kainuu - Outokumpu, 2 - Onegozersko - Rybinsk, 3 - Vetreny Payas, 4 - Lapland, 5 - Pechenga - Varzuga.

$PR_1^3$. 4 - axial zones of mafic-ultramafic magmatism: 1 - Dalsland (revised by author), 2 - Kylmakoski, 3 - North Swedish, Lapland - White Sea suture); 5 - Archean crust segments intensively reworked in Svecofennian time; 6 - terrains of alkaline - ultramafic series (Gremyakha - Vyrmes, Yeletozero - Tiksheozero); 7 - boundary between the Baltic Shield and the Timanides and Caledonides.

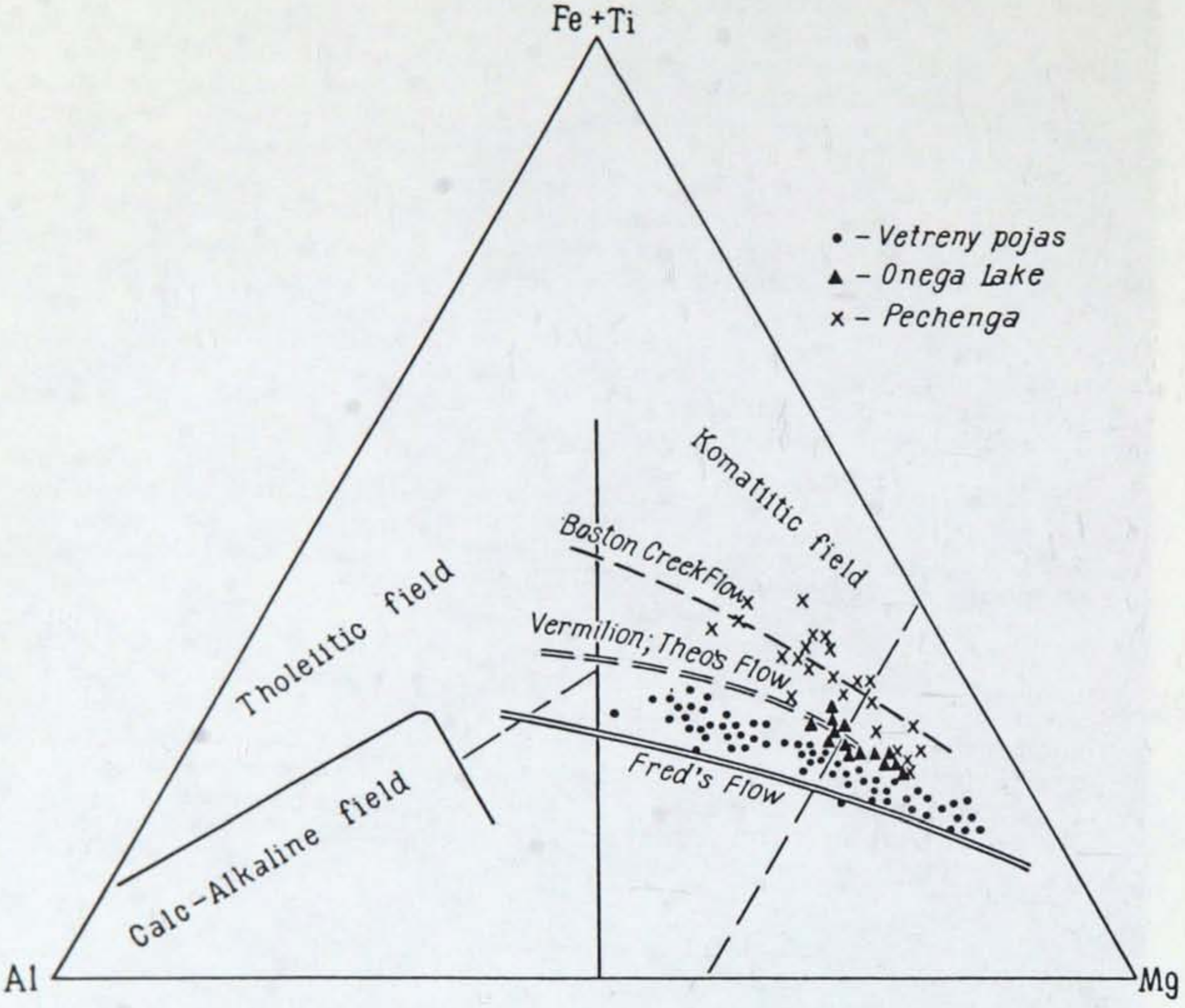


Fig. 2. Jensen cation plot for Early Proterozoic (Ludicovian) high-magnesian rocks of the tholeiite series (Lake Onega and Pechenga) and the komatiite series (Vetreny Poyas), Baltic Shield. The high-magnesian series of the Canadian Shield after Stone et al. (1987).

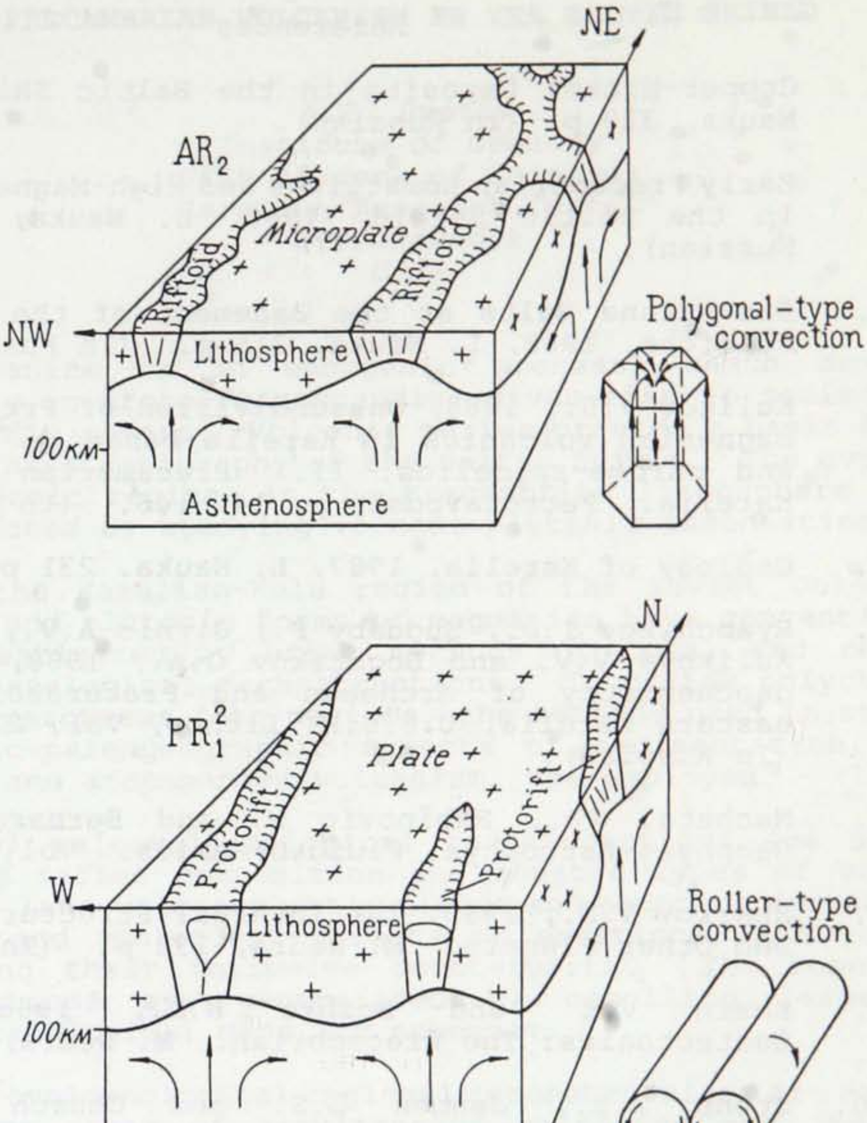


Fig. 3. Hypothetical types of convection cells in the Late Archean ($AR_2$) and Early Proterozoic (Ludicovian, $PR_1^2$) asthenosphere of the Baltic segment.

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# PRECAMBRIAN VOLCANISM IN THE BALTIC SHIELD

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Volcanism as an endogenic process, which developed regionally in proto-Fennoscandia, gives rise to sedimentary-volcanogenic covers. Volcanic series provide a basis for the Precambrian stratigraphy of the Baltic Shield. The evolution of geodynamic regimes in the Precambrian lithosphere can be reconstructed by studying volcano-plutonic associations.

In the Karelian-Kola region of the Soviet Union, the volcanic and plutonic forms of magmatism have conventionally been investigated by using methods of local and regional paleovolcanological reconstructions. To divide polychronous volcanic paroxysms into periods, the methods used in studying lithologic-paleogeographic aspects of sedimentation, which preceded and accompanied volcanism, are employed.

Local paleovolcanological reconstructions are used to study the facies composition and genetic types of volcanic products, lava-volcanoclastic fields and eruption centers, the dynamics and geological effects of eruptions, and volcanic series and their intrusive counterparts. The results of investigations are generalized by compiling large-scale paleovolcanological maps and schemes.

Paleovolcanological regional reconstructions are employed to restore areas of simultaneous volcanism and its total geological effect, to define the centers of inherited endogenic activity and ore formation and the paleotectonic and paleogeographic settings of volcanic epochs, as well as to provide a basis for paleovolcanologic demarcation. The results obtained are shown on small-scale paleovolcanological schemes for a series of age sections. Local and regional paleovolcanological reconstructions are made to develop volumetric models of volcano-plutonism which are observed as individual centers of endogenic activity and as large provinces.

Igneous rocks of volcano-plutonic series and associated sedimentary-volcanogenic strata show some laterally persistent structural, textural and compositional features used as criteria for their regional correlation. A series of lithologic-paleogeographic and paleovolcanological studies provides a basis for elaborating the Precambrian regional stratigraphy of the Baltic Shield (Fig. 1).

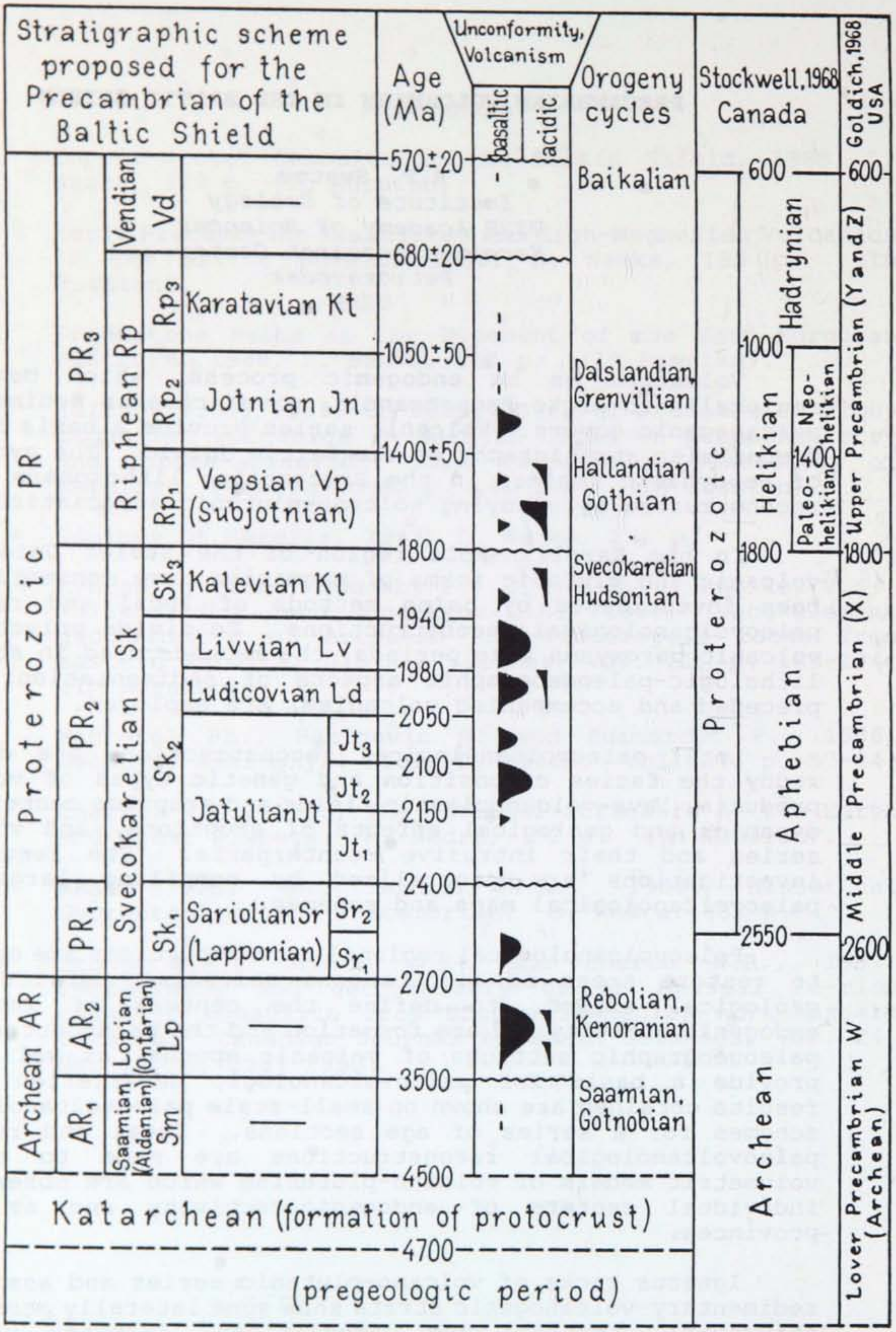



Fig. 1. Chronostratigraphic correlation scheme for the Precambrian of the Baltic Shield and the Canadian Shield (based on the author's data; Geology of the Karelian, 1987; Goldich, 1968; Keller, 1979; Lundqvist, 1979; Merilainen, 1980; Salop, 1982; Stockwell, 1968; Stratigraphy Precambrian of the Karelian ASSR, 1984).

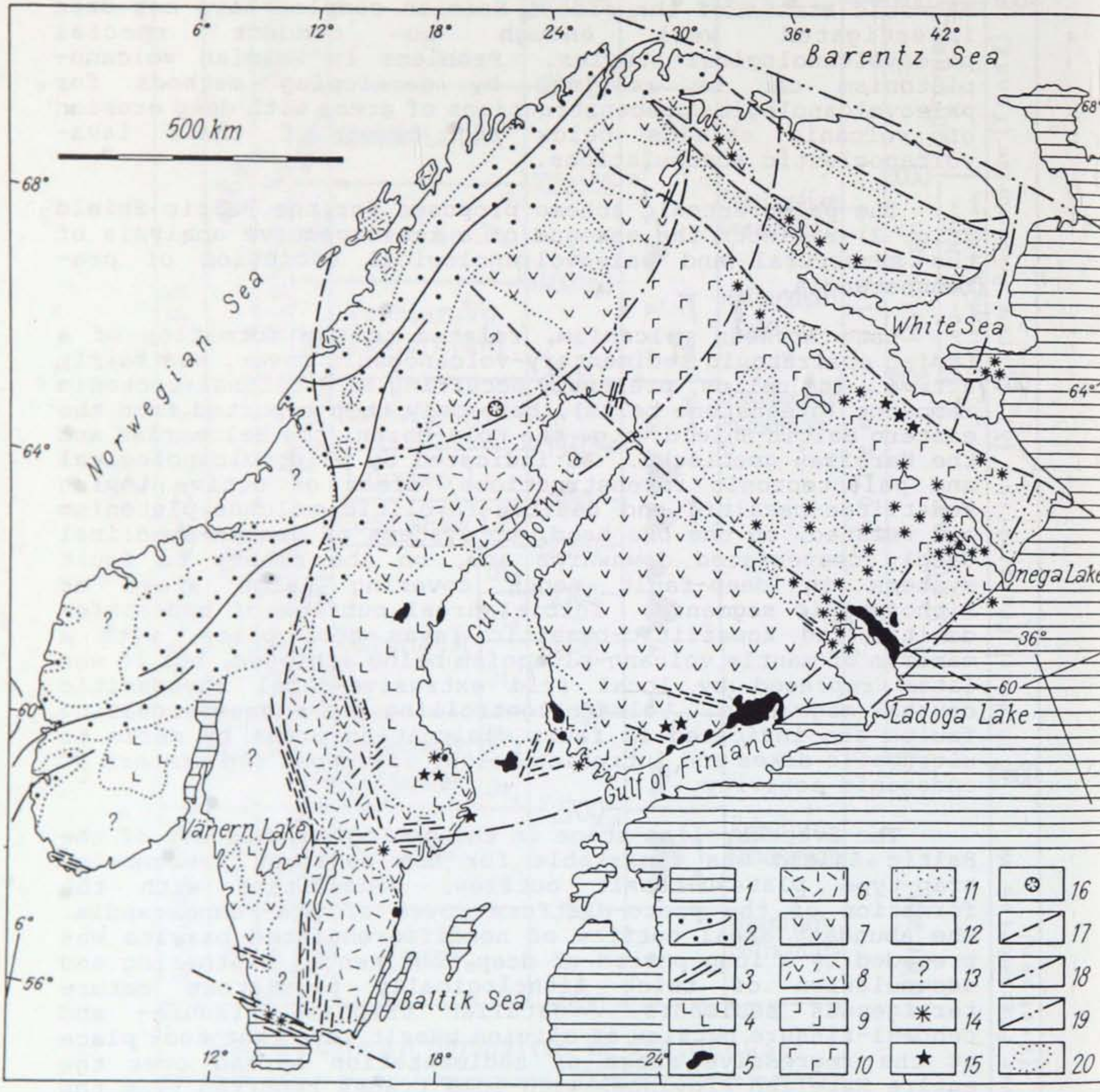
Paleovolcanological reconstructions of the earliest events in the Precambrian volcanic record can be effectively made from the Late Precambrian onwards. The early Archean magmatic strata of the oldest Saamian complex have not been investigated well enough to conduct special paleovolcanological studies. Problems in Saamian volcano-plutonism can be resolved by developing methods for paleovolcanological reconstructions of areas with deep erosion of volcanic shields below the bases of their lava-volcanoclastic accumulations.

The paleotectonic scheme proposed for the Baltic Shield (Fig. 2) reflects the essence of a retrospective analysis of the structural and paleovolcanological evolution of pra-Fennoscandia.

Late Archean volcanism, related to the formation of a Lopian epicratonic sedimentary-volcanogenic cover, was fairly active. Its relics, presently occurring as erosional-tectonic remnants (greenstone belts), have only been reported from the eastern Baltic Shield, e.g. the Kola-Mezen, the Belomorian and the Karelian geoblocks. As indicated by paleovolcanological and paleotectonic reconstructions, areas of active Lopian komatiitic-basaltic and dacitic-rhyolitic volcano-plutonism are related, on the one hand, to systems of graben-synclinal deeply compensated downwarps and, on the other, to fault systems in deep-fault zones covering large areas of lithospheric segments. Initial areal outflow of andesitic-dacitic and komatiitic-basaltic lavas took place, with a maximum of mantle volcano-plutonism being achieved, but it was later replaced by local acid extrusive-domal rhyodacitic crustal magmatism. Volcano-controlling and magma-evacuating faults are indicated in fault dislocation areas by mafic to ultramafic dikes and intrusion belts, volcanoes and centers of endogenic activity.

The Svecokarelian stage in the geological history of the Baltic Shield was remarkable for the earliest Precambrian trap-type plateau-basalt outflow, concomitant with the formation of the proto-platform cover of pra-Fennoscandia. The abundant areal outflow of nondifferentiated basalts was preceded by a long period of deep continental weathering and accumulation of thick lithologically persistent mature terrigenous sediments. Jatulian triphase fissure- and central-fissure outflow of olivine basalt lava that took place at the regressive stage of sedimentation spread over the entire Karelian Province, with some traces reported from the Svecofennian Province. Volcanic series of plateau basalts are characterized by the presence of ferro- and cuprobasalts in both lava and subvolcanic facies.

The lava-volcanoclastic fields of Ludicovian plateau basalts occupy the Svecofennian Province. In the Karelian Province they occur mainly in volcano-tectonic structures


6°
12°
18°
24°
30°
36°
42°
Barentz Sea
68°
500 km
68°
Norwegian Sea
White Sea
64°
64°
Gulf of Bothnia
Onega Lake
36°
60
60
Ladoga Lake
Gulf of Finland
Vänern Lake
6°
56°
Baltik Sea
12°
18°
24°
1
2
3
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5
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Fig. 2. Paleotectonic scheme of the Baltic Shield (made by the author and based on the data published by the Geological Surveys of the USSR, Finland, Sweden and Norway).

1 - East European platform cover

2 - Caledonian geosyncline and its axial lines

3 - undivided dike belts and fields

4 - reconstructed areas of Riphean volcanism

5 - rapakivi granite terrains

6 - Trans-Scandinavian granite porphyry belt

7 - undivided Svecokarelian plateau basalt provinces

8 - undivided Svecokarelian provinces of plateau basalt and rhyodacite volcanism

9 - Livvian high-magnesian basalt fields

10 - Jatulian plateau basalt fields

11 - undivided areas of Lopian volcanism and plutonism

12 - Early Archean protocrustal areas (central axis)

13 - undivided areas of newly-formed crust

14 - volcanoes resulting from basaltic volcanism

15 - extrusive domes and centers of acid volcano-plutonism

16 - caldera

17 - major tectonic geoblock-linkage zones

18 - axial lines of regional radial flexures

19 - tectonic shield-limiting zones (Karpinsky line; Tornqvist-Teisseyre zone)

20 - generalized boundaries of volcanic fields, areas and provinces.

regarded as centers of endogenic activity. Areas of Jatulian-Ludicovian volcano-plutonism are formed by the merged bases of Iceland-type shield volcanoes. The Ludicovian mantle plateau basalt fields in the Svecofennian Province are conjugated with areas of shallow crustal extrusive- domal rhyodacitic volcano-plutonism, widespread in Sweden and less common in southwest Finland. It was not until endogenic activity reached its maximum in the Livvian that high-magnesian volcano-plutonism began to occur locally as picritic basalt fields and tuffs and shallow peridotite intrusion belts with Cu-Ni metallization in plateau basalt areas. The eruption centers of high-magnesian magmatites are usually coupled with fault systems. The latter resulted from the formation of lithospheric geoblock suture zones affected by deep-fault dislocations in the extensional regime.

The cessation of fading Svecokarelian volcano-plutonism was marked by local occurrences of Kalevian andesitic-basaltic and basaltic volcano-plutonism, the eruption centers of which inherited areas of preceding Jatulian-Livvian mantle volcanism.

The total volume of Svecokarelian volcano-plutonism exceeds 350,000 $km^3$ and the total area of evacuated basalts is as much as 0.5 million $km^2$.

Early Riphean (sub-Jotnian) bimodal volcano-plutonism was most active in southwest Sweden along the Protogin zone which hosts both the Smaland and the Varmland granite-rhyodacitic extrusive-domal plateaus that form part of the structurally localized linear Trans-Scandinavian granite-porphyry belt. The role of basaltic volcano-plutonism in the above region is minor. Also, it has been reported locally from the Karelian Province. This period is also referred to as a "rapakivi granite epoch", which is remarkable for gabbro-anorthosite and rapakivi granite intrusions as well as local rhyodacite intrusions.

Acid volcano-plutonism had reached its maximum by the end of the Early Riphean, but later its activity gradually decreased to die out completely.

Late Riphean volcanism, concomitant with the formation of the platform sedimentary cover, is indicated by evacuated olivine basalts (plateau basalts) in Central Sweden and local outflow in the pericratonal grabens that make up the Baltic Shield margins in the marginal radial flexure zone. The outflow of plateau basalts was accompanied by the formation of layered gabbro-norite and gabbro-dolerite sills and by a few dolerite dike intrusions. Precambrian volcanism ceased in Vendian time.

Subsequent Phanerozoic volcanic activity is associated with the formation of a trans-structural Caledonian geosynclinal belt. It also occurred as eruptions of short duration along the margins of the Baltic Shield due to its isostatic uplift; these continued to the Oligocene.

The regime in which regional flexures developed played an important role in the areal distribution of both Precambrian and Phanerozoic volcano-plutonism. The flexures were transformed later in the course of radial and deep-fault tangential dislocations. As a result, en echelon-faults, partly controlling the locations of volcanics and magma-feeder systems, were formed.

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# LATE ARCHEAN SEDIMENTARY AND VOLCANOGENIC DEPOSITS IN THE BOUNDARY ZONE BETWEEN THE KARELIAN AND BELOMORIAN SEGMENTS OF THE BALTIC SHIELD IN THE LAKE KERET AREA, NORTHERN KARELIA (ON THE RELATION OF THE LOPIAN AND BELOMORIAN SUPRA-CRUSTAL COMPLEXES)

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The Karelian and Belomorian segments of the Earth's crust are major tectonic elements in the northeastern Baltic Shield (Fig. 1A). The former is identified as a granite-greenstone area and the latter as a highly metamorphosed area.

Geophysically, the above areas are fairly distinct. Thus, it can be seen from the scheme for the Earth's crustal thickness in Karelia (Fig. 1C), based on seismic and gravimetric data, that the estimated depth of the M-discontinuity is less in the Belomorian segment than in the Karelian segment.

Relations between the Earth's crustal segments represent one of the most disputable and widely discussed problems in early Precambrian geology. In this connection, relations between the supracrustal complexes characteristic of the above segments (e.g., the Lopian complex in the Karelian area and the Belomorian complex in the Belomorian area) are of primary importance.

The Lopian complex is built by the late Archean (2.7-3.2 Ga) sedimentary and volcanogenic deposits that form part of the greenstone belts in the Karelian granite-greenstone area.

The Belomorian complex consists of polymetamorphic deposits subdivided into three suites (10):

1) the Keret suite which is composed of gneissose granites comparable with "grey gneisses";

2) the Hetolambina suite composed of amphibolites with gneiss streaks; and

3) the Chupa suite represented by alumina gneisses with amphibolite streaks.

The rocks of the Belomorian complex have suffered high-pressure (kyanitic type) moderate-temperature metamorphism

(1).

There are three points of view regarding the relations of such complexes (6):

1. Highly and poorly metamorphosed complexes differ in both age and the structural environment in which they were formed.

2. Highly metamorphosed deposits represent deeply eroded counterparts of granite-greenstone associations.

3. The above complexes generally show the same age and are used as markers for different structural-formational zones.

Similarly, the geologists who study the Karelian region have differing viewpoints on the relationships of these tectonic elements.

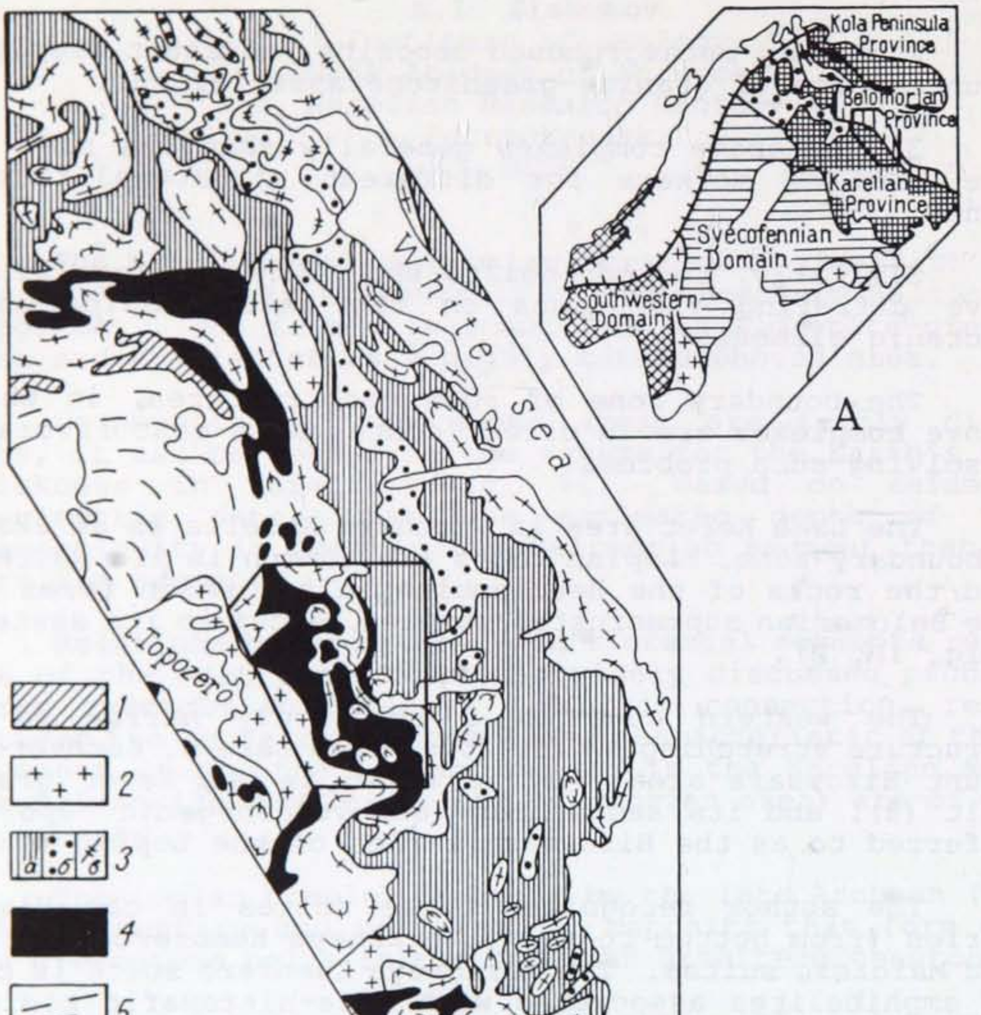
The boundary zone of such geostructures, in which the above complexes are in direct contact, is most favorable for resolving such problems.

The Lake Keret area in northern Karelia is an example of a boundary zone. Lopian rocks are common in its western part and the rocks of the Hetolambina suite, which forms part of the Belomorian supracrustal complex, occur in its eastern part (Fig. 1A, 2).

The western zone is a relatively narrow (1-15 km) structure stretching NW for 80 km from Lake N. Kumozero to the Mount Hizovaara area. It is known as the Keret greenstone belt (8), and its sedimentary and volcanogenic deposits are referred to as the Hizovaara series of the Lopian complex.

The author recognizes three suites in the Hizovaara, series (from bottom to top): Verkhneye Kumozero, Hattomozero and Maiozero suites. The Verkhneye Kumozero suite is composed of amphibolites associated with meta-ultramafic bodies. As indicated by petrochemical and textural characteristics, the primary nature of the amphibolites corresponds to basalts (11) and that of the meta-ultramafic rocks to komatiites. In the Mount Hizovaara area the volcanogenic nature of the above meta-ultramafic rocks has been proved by distinguishing individual flows separated by tuffaceous interbeds, each clearly differentiated (5).

In the eastern part of the structure in the Lake Verkhneye-Shobozero area, the base of the suite is represented by a (kyanite)-garnet-biotite gneiss member a few tens of meters in thickness which is probably a paramember. The rocks of the suite are about 400 m thick.


Kola Peninsula Province
Belomorian Province
Karelian Province
Svecofennian Domain
Southwestern Domain
A
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L.Topozero
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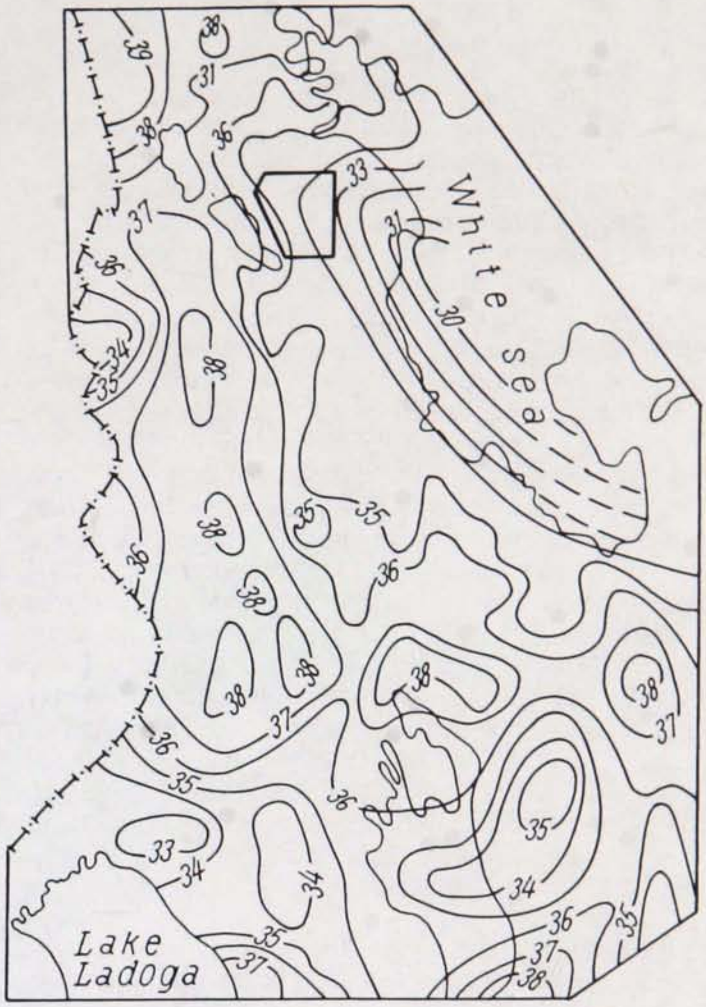
20 0 20km
B

Fig. 1




C

Fig. 1 A. Tectonic demarcation of the Baltic Shield (Gaal et al., 1989) and the location of the region discussed.

Fig. 1 B. Geological structure of the NW White Sea region, by A.I. Slabunov. Based on publications of the Karelian Research Center and Gorlov, 1967; Markov et al., 1987; Stepanov & Slabunov, 1989. 1 - Karelian ($PR_1$) supracrustal complexes; 2 - Early Proterozoic intrusive charnockites and enderbites; 3 - Belomorian supracrustal complex: a - amphibolites alternating with gneisses (Hetolambina suite); b - "alumina" gneisses (Chupa suite and its counterparts); c - biotitic and amphibole-biotite gneisses; 4 - Lopian ($AR_2$) supracrustal complex; 5 - locally microclinized plagiogranites, diorites and quartz diorites (Tavajarvi complex); 6 - heterogeneous granite gneiss ("grey gneiss") complex.

Fig. 1 C. Map of the Earth's crustal thickness in the southeastern Baltic Shield (after A.S. Grishin, in Geology of Karelia, 1987).

The overlying Hattomozero suite is composed of epidote-amphibole-biotitic and biotite gneisses which, as a rule, retain relics of agglomeratic texture. This feature, as well as their petrochemical characteristics, provide a basis for identifying the above rocks as intermediate metatuffs. Also, two volcanoes consisting of tuffs have been reconstructed. The center of one of them is in the Mount Hizovaara area (9) and that of the other is in the Lake Severnoye-Hattomozero area. In the Mount Hizovaara area the suite is more varied in composition, with metasediments and subvolcanic rhyolite bodies (9). The suite varies in thickness from 700 to 1300 m.

The Maiozero suite represented by amphibolites with paragneiss interbeds resting unconformably on the underlying rocks. The unconformity has been unambiguously established in the Mount Hizovaara area (9). Meta-komatiite bodies have been reported from the lower part of this sequence in the above area (5). Relics of pillow textures are often preserved in amphibolites. This fact together with their petrogeochemical characteristics indicate that the rocks were metabasalts and help to determine the tops and bottoms of sequences. The suite is about 300 m thick.

The thickness of the entire supracrustal complex within the Keret greenstone belt represented by the Hizovaara series varies from 1200 to 2100 m.

The Hizovaara series in the linear-folded (western) zone is typical of late Archean (Lopian) greenstone belts in Karelia (2,3). They are similar in both rock constituents and in their stratigraphy.

The eastern zone is a vast field in which gneissose granite domes and arches are fairly common among the rocks of the Belomorian supracrustal complex. It is clearly seen on the geological scheme (Fig. 2) that the gneissose granites occupy the cores of antiforms. The gneissose granites represent autochthonous synmetamorphic deposits which developed during the granitization of the sialic basement which is preserved as tonalite gneiss relics.

The eastern (gneiss-domal) zone is separated from the greenstone belt by a gneissose granite field varying in width from 200 m to 10-20 km (Fig. 2). In the westernmost part of the zone, adjacent to the greenstone belt, the supracrustal complex consists of three units: a lower unit composed of amphibolites, a middle unit of metatuffs (including agglomeratic metatuffs), and an upper unit formed by amphibolites with (kyanite)-garnet-biotite gneisses. The above deposits are similar to the Hizovaara series: the lower unit is represented by the Verkhneye Kumozero suite, the middle unit by the Hattomozero suite, and the upper one by the Maiozero suite. A few kilometers to the east, the supracrustal complex is also similar to the Hizovaara series, but the Verkhneye Kumozero suite is much thinner (ca 10 m).

In the northwestern part of the gneiss-domal zone, the supracrustal complex is subdivided into two units: a lower unit composed of amphibolites with garnet-biotite gneiss interbeds and an upper unit which consists of alumina gneisses with magnetite-bearing garnetite intercalations. The lower unit, if traced laterally southwestward, merges with the Maiozero suite (Fig. 2) and can thus be unambiguously identified. The supracrustal rocks of the gneiss-domal zone, comparable with the Hizovaara series, vary in thickness from 700 to 1000 m and tend to thin east of the greenstone belt because the two lower suites wedge out.

The Maiozero suite, reported from both zones, is a marker unit which can be traced from the greenstone belt to the gneiss-domal zone.

The data presented make it possible to correlate the Lopian and Belomorian supracrustal complexes in the Lake Keret area. These deposits are the same age but developed in different structural-formational zones of a single late Archean granite-greenstone system.

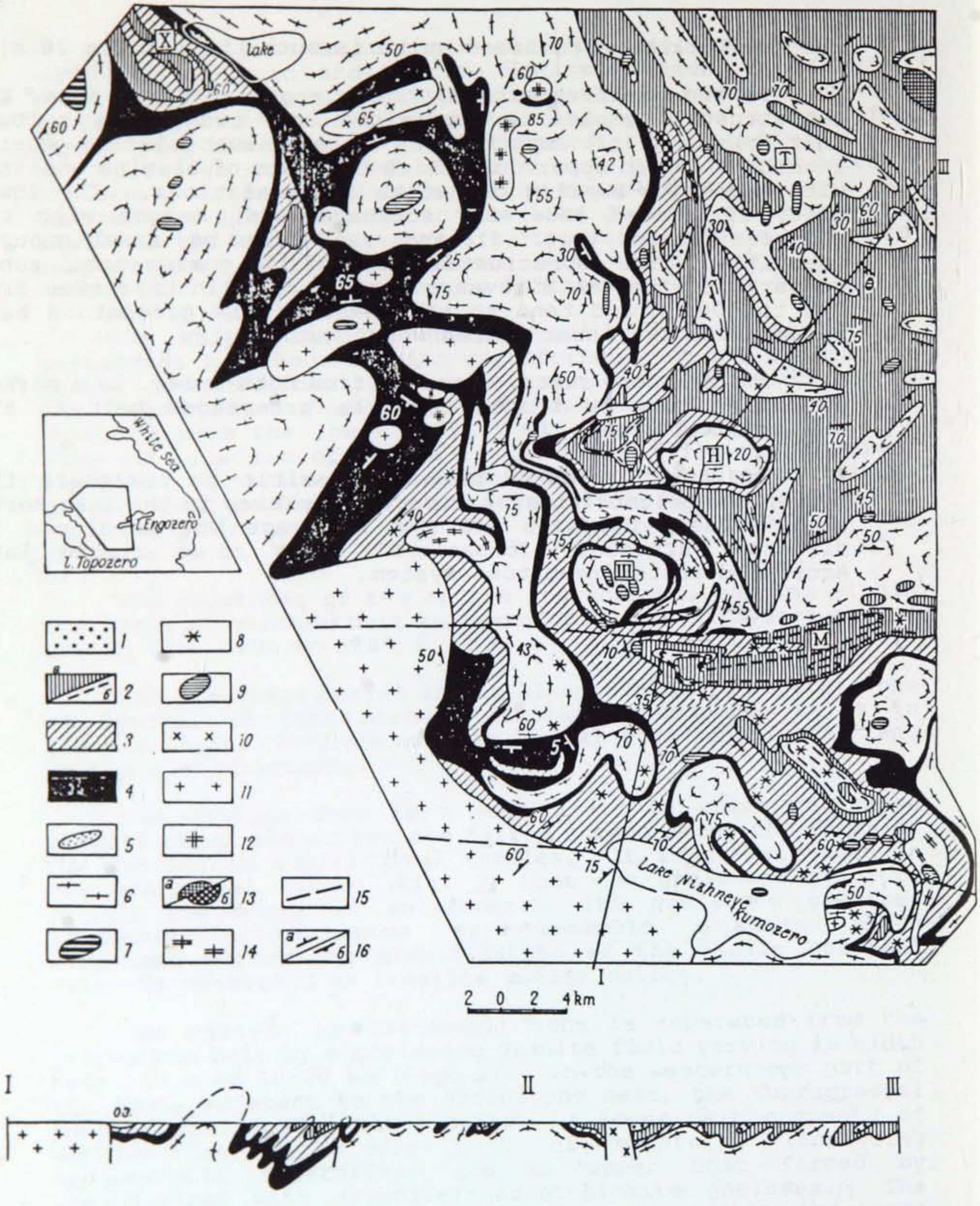

Lake
White Sea
L. Engozero
L. Topozero
Lake Nizhneye Kumozero
2 0 2 4 km
I
II
III

Fig. 2. Geological structure of the Lake Keret area by A.I. Slabunov. Based on data obtained by A.I. Slabunov, O.I. Volodichev, V.N. Kozhevnikov, V.S. Stepanov, J.I. Systra, V.V. Shchiptsov, V.V. Yuzhanova, N.A. Volotovskaya and E.P. Chuikina.

1 - "alumina" gneisses with magnetite-enriched garnetite interbeds; 2-4 - suites of the Hizovaara series: 2 - Maiozero suite: a) amphibolites with (kyanite)- garnet-biotite gneisses, b) mappable (kyanite)-garnet-biotite gneiss interbeds; 3 - Hattomozero suite; 4 - Verkhneye Kumozero suite; 5 - garnet- biotite gneisses in the Lake Verkhneye-Shobozero area; 6 - gneissose granites; 7 - Logijarvi ultramafic intrusion; 8 - garnet gabbro; 9 - intrusions of the lherzolite-gabbro norite complex; 10 - plagiomicrocline granites; 11 - metacharnockites; 12 - metaenderbites; 13 - intrusions of the gabbro-anorthosite complex: a) with relics of primary structure, b) intensely amphibolitized; 14 - allochthonous microcline-bearing plagiogranite gneisses; 15 - hypothetic fault zones; 16 - planar elements of occurrence: a - banding, b - gneissosity.

X - Hizovaara structure, M - Maiozero synform; domal structures: W - Shobozero structure (Lakes Verkhneye and Nizhneye Shobozero), H - Lake Pervoye Nogtevo-Paiozero structure, T - Lake Travyanoye structure.

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# EARLY PROTEROZOIC SEDIMENTARY BASINS OF THE BALTIC SHIELD

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A basis for modern evolutionary models of the Baltic Shield is mainly provided by geochemical and tectonic data. This paper is an attempt to use information on the lithology of Early Proterozoic formations for analysis and to reveal the geodynamic characteristics of the region paleogeographically. Data on the relations of sedimentary, volcanic and intrusive formations (Table 1) allow for lithologic and geochronologic correlation of Proterozoic stratigraphy between the eastern and western parts of the Baltic Shield.

Sumian deposits of the eastern part of the shield accumulated in a continental volcanic arc that was initiated by the collision and subduction of Belomorides rocks beneath the Karelian block that was a product of Archean cratonization. The result was a narrow extended basin of coarse clastic sedimentation and abundant acidic volcanism.

In Sariolian time, the area of active sedimentation became much wider (Fig. 1). It consisted of grabens on arched uplifts; shallow basins connected the grabens. Extension occurred along the axes of the basins. The source area of the sediments roughly coincided with the Archean Belomorian block. Coarse-grained volcaniclastics and sediments accumulated in the basin along with andesitic-basaltic, basaltic and locally komatiitic lavas.

At the beginning of Jatulian time, the shield was a peneplain although the nature of its southwestern part is unknown (Fig. 2). Active chemical weathering supplied mature clastic material. Spreading of the crust along both branches of the extensive shallow basin resulted in linear sags compensated by thick (2-4 km) volcanics and coarse clastics. The pelitic component of weathered material accumulated in the northern part of the basin forming a thick shale unit.

By the end of Jatulian time the region was almost entirely occupied by the shallow basin with clastic-carbonate sedimentation locally interrupted by extrusion of tholeiitic basalts. In the northeast an oceanic structure was possibly generated simultaneously with activation of its northeastern border where dacitic-rhyolitic and basaltic volcanism occurred.

The Ludicovian period began with the development of a classical geodynamic system, including a spreading zone, a young ocean, island and continental volcanic arcs, a rift, and back-arc basins (Fig. 3). These tectonic elements are based on the presence of ophiolites, olistostromes of initial rifting, coarse clastics and turbidites of the continental slope, hydrothermal accumulations along the axial spreading zone, sediments and volcanics of magmatic arcs, komatiites of back-arc areas and geochemical characteristics. The nature of some greenstone belts and some (subisometric) structures of that time remain obscure.

Kalevian time corresponds to the closure of the Ludicovian ocean and the accumulation of flysch sediments in the vicinity of the suture. The presence of a volcanic arc in the south is assumed.

The formation of new crust by the accumulation of oceanic turbidites, arc volcanics and other sediments was dominant during Svecofennian time (Fig. 4) as the volcanics and sediments were accreted onto the craton.

Isolated Vepsian continental basins were filled with terrigenous redbeds.

In general, the Early Proterozoic history of the shield consists of two defineable Wilson cycles which resulted in the accretion of the southwestern part of the shield.

One of the distinctive features of Proterozoic sedimentation was the essential role of terrigenous material through all the periods. It was present in large quantities even in oceanic sediments. The typical compensated character of the basins can probably be explained by the comparative mobility of clastic material in an environment that had no land vegetation and had abundant precipitation.

The climate in Sumian and Sariolian time was cold, as indicated by the absence of weathering and the presence of diamictites, dropstones and perhaps tills. The presence of a weathering crust, caliche, and mature sediments are indicative of a warm and humid but sometimes arid environment during the following Jatulian period. Signs of aridity became more pronounced with the presence of evaporites in the Ludicovian and later periods.

Diverse types of mineralization and ores are characteristic of different deposits depending on their geodynamic nature.

The Early Proterozoic history of the Baltic Shield is very similar to the history of other Precambrian regions.

Table 1. Summary of Early Proterozoic Stratigraphy of the Baltic Shield

| Supergroup (Superhorizon) | Age Ga | Typical deposits | | Typical formations |
|---|---|---|---|---|
| VEPSIAN | 1.75 | quartzite, conglomerate | granite (1.79-1.75) | Vakko, Rissavaara Kumpu, Dobblon (1.80) |
| | | | unconformity | |
| SVECOFENNIAN | $1.85 \pm 0.01$ | rhyolite, andesite basalt, conglomerate | granodiorite, granite (1.88-1.85) | Kiruna porphyry (1.89-1.86) Arvidsjar (1.88) |
| | | | unconformity | |
| KALEVIAN | $1.92 \pm 0.01$ | schist, quartzite, conglomerate | Haparanda granite (1.89-1.86) Ern (1.89) | Kurravaara, Pahakurkio |
| | | | unconformity | |
| LUDICOVIAN | $1.95 \pm 0.01$ | black schist, dolomite, iron-formation chaotic breccia | albite diabase (22.2-1.95) | Kiruna greenstones Kolari greenstones |
| | | | unconformity | |
| JATULIAN | $2.08 \pm 0.02$ | dolomite, schist basalt, sandstone, quartzite, conglomerate | gabbro-wehrlite association (2.2-2.1) | quartzitic fms |
| | | | unconformity | |
| SARIOLIAN | $2.30 \pm 0.05$ | conglomerate, agglomerate, andesite-basalt (2.33), basalt | gabbro | Goldenvarry, Holmvatn |
| | | | unconformity | |
| SUMIAN | $2.40 \pm 0.05$ | quartz porphyry, arkose, conglomerate | | |
| ARCHEAN | $2.60 \pm 0.10$ | | | |

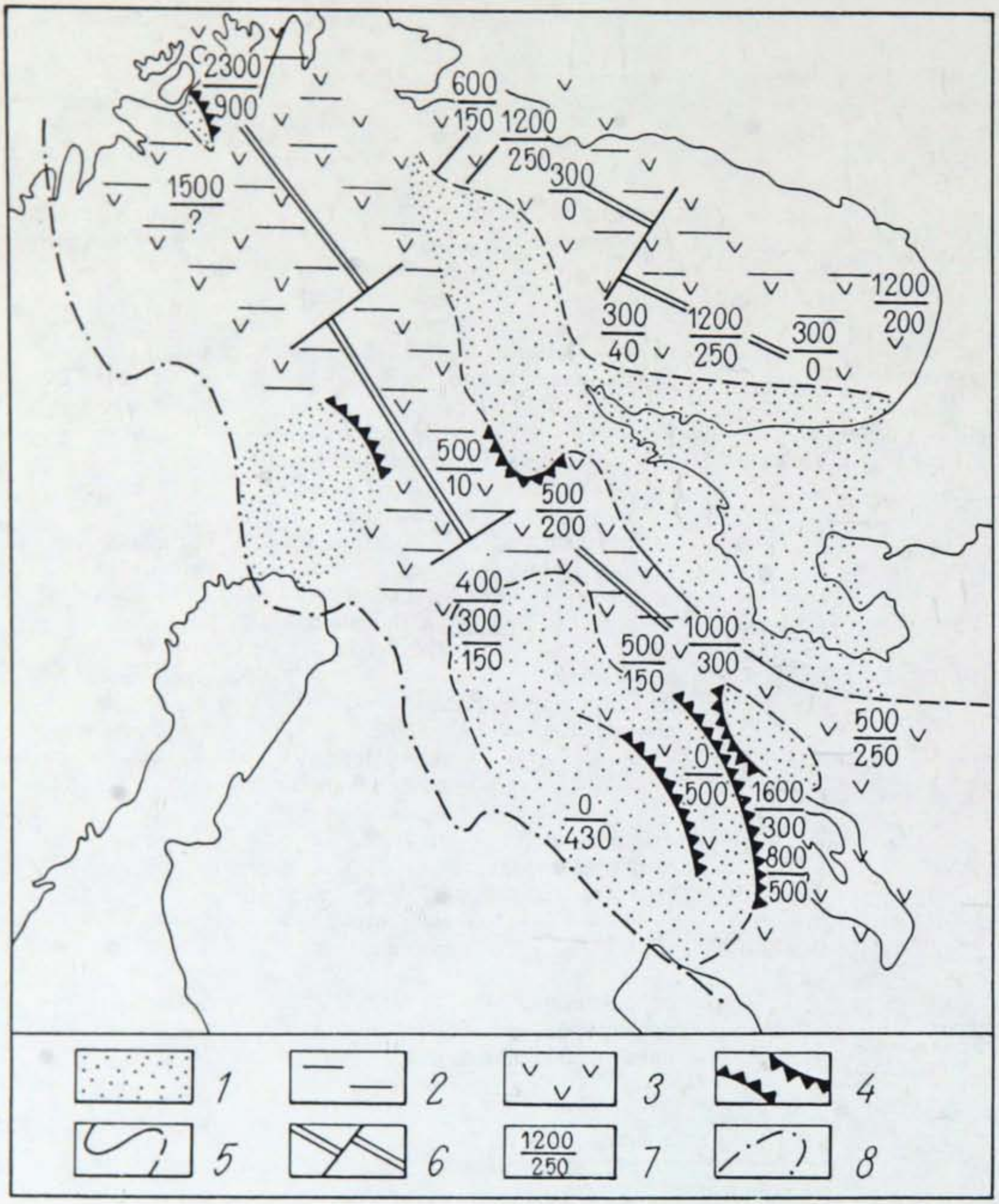


Fig. 1. Sariolian sedimentation and tectonics in the Baltic Shield.

1 - area of clastic sedimentation; 2 - sedimentary basin; 3 - area of mafic volcanism; 4 - boundary graben; 5 - boundary of sedimentary areas, established and assumed; 6 - stretching axes and transform faults; 7 - thickness of volcanics (above) and terrigenous rocks (below); 8 - approximate southwestern boundary of Karelian craton.

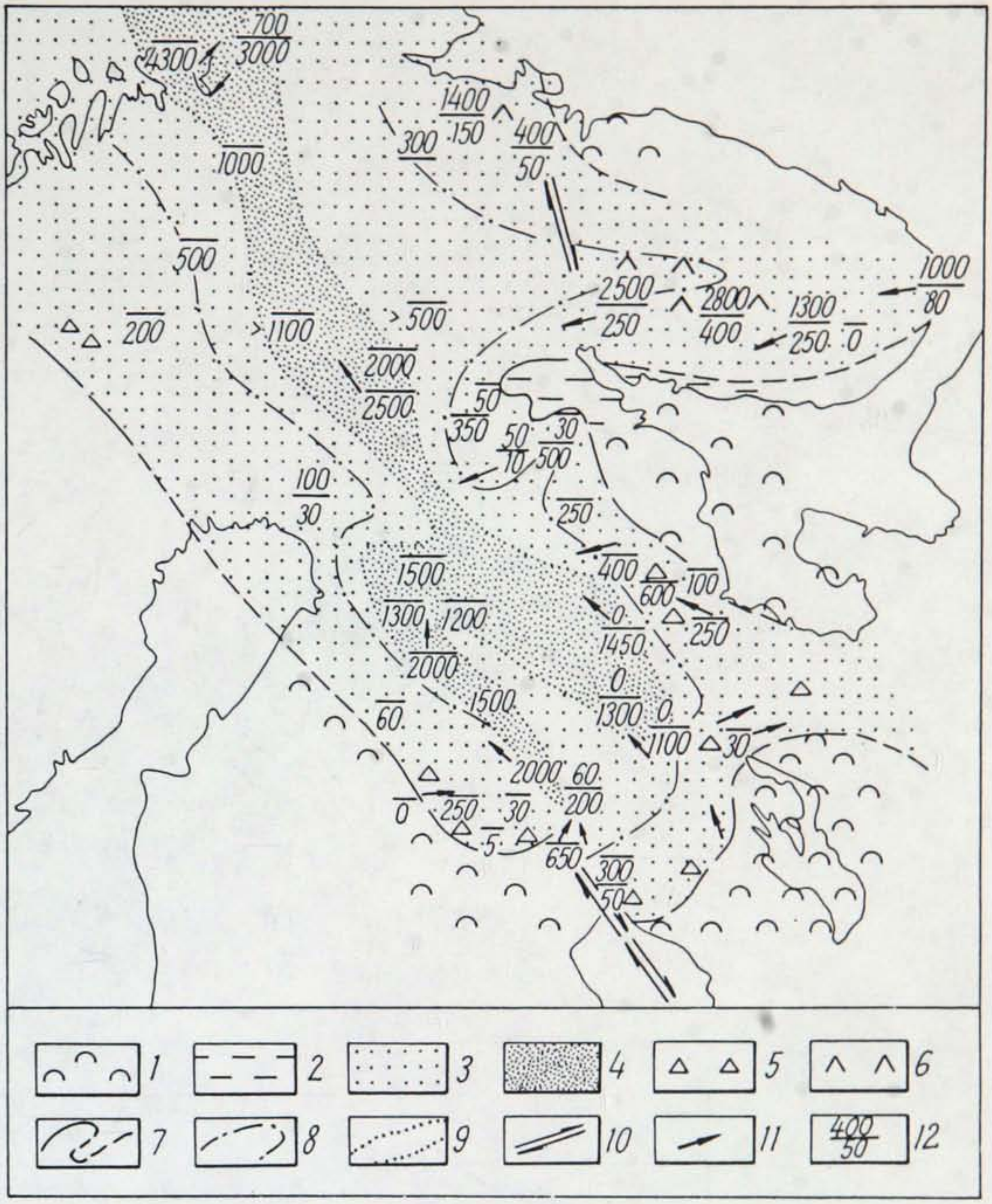


Fig. 2. Early Jatulian sedimentary basin.

1 - stable source areas, with hilly relief; 2 - same as 1, lowland; 3 - terrigenous sediments up to 1000 m in thickness; 4 - terrigenous sediments more than 1000 m thick; 5 - carbonatized eluvium; 6 - acidic volcanism; 7 - area of accumulation; 8 - 500 m isopach line; 9 - 1000 m isopach line; 10 - post-Jatulian fault; 11 - direction of sediment transport; 12 - thickness of volcanics (above) and sediments (below).

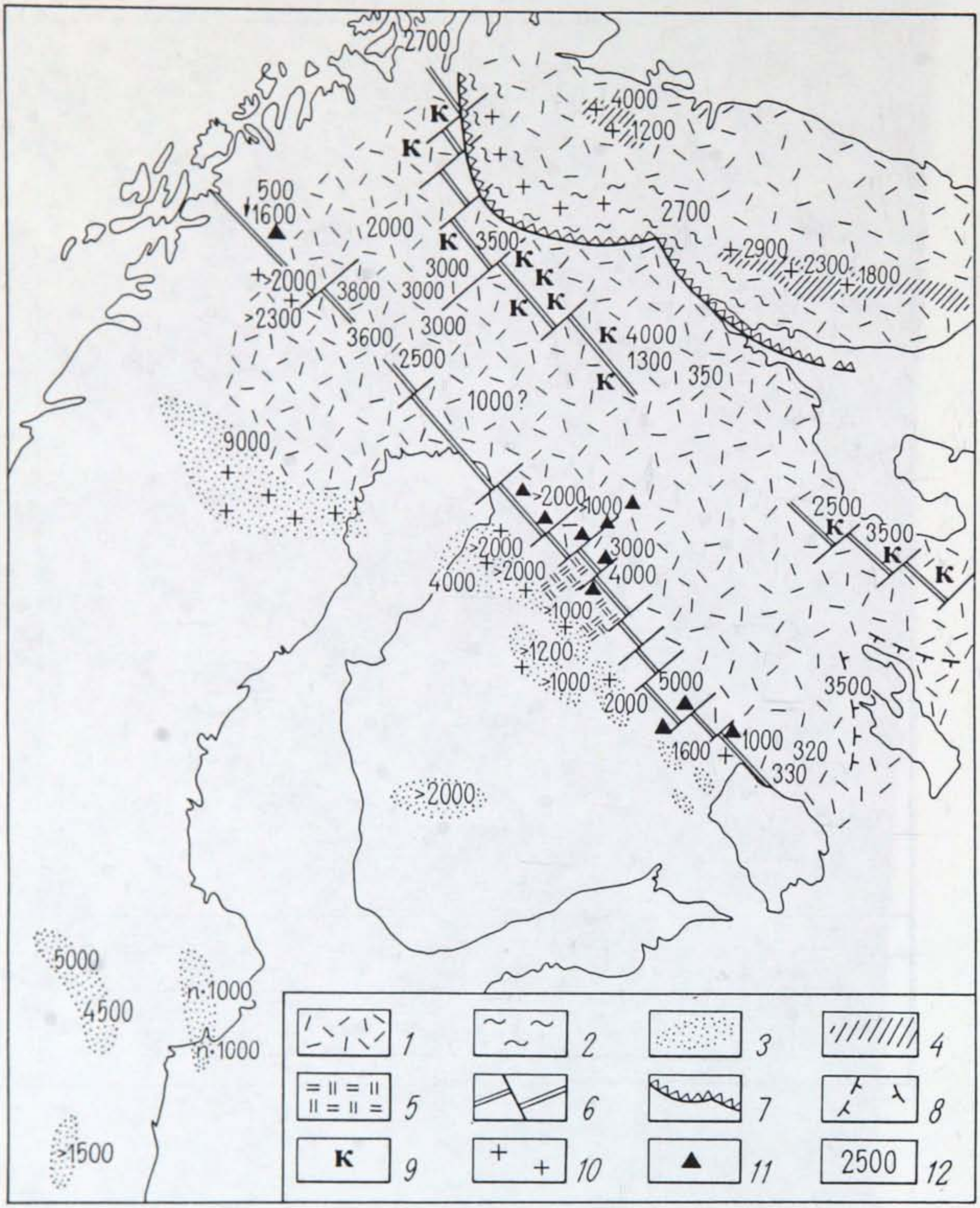


Fig. 3. Ludicovian geodynamic system.

1 - Archaean cratonized area; 2 - granulite belt; 3 - pre-Svecofennian blocks in Svecofennides; 4 - Pechenga - Varzuga belt; 5 - Iisalmi microcontinent; 6 - spreading axes and transform faults; 7 - subduction zone; 8 - aulacogen; 9 - komatiites; 10 - acidic volcanics; 11 - olistostromes and coarse sediments; 12 - thickness of deposit.

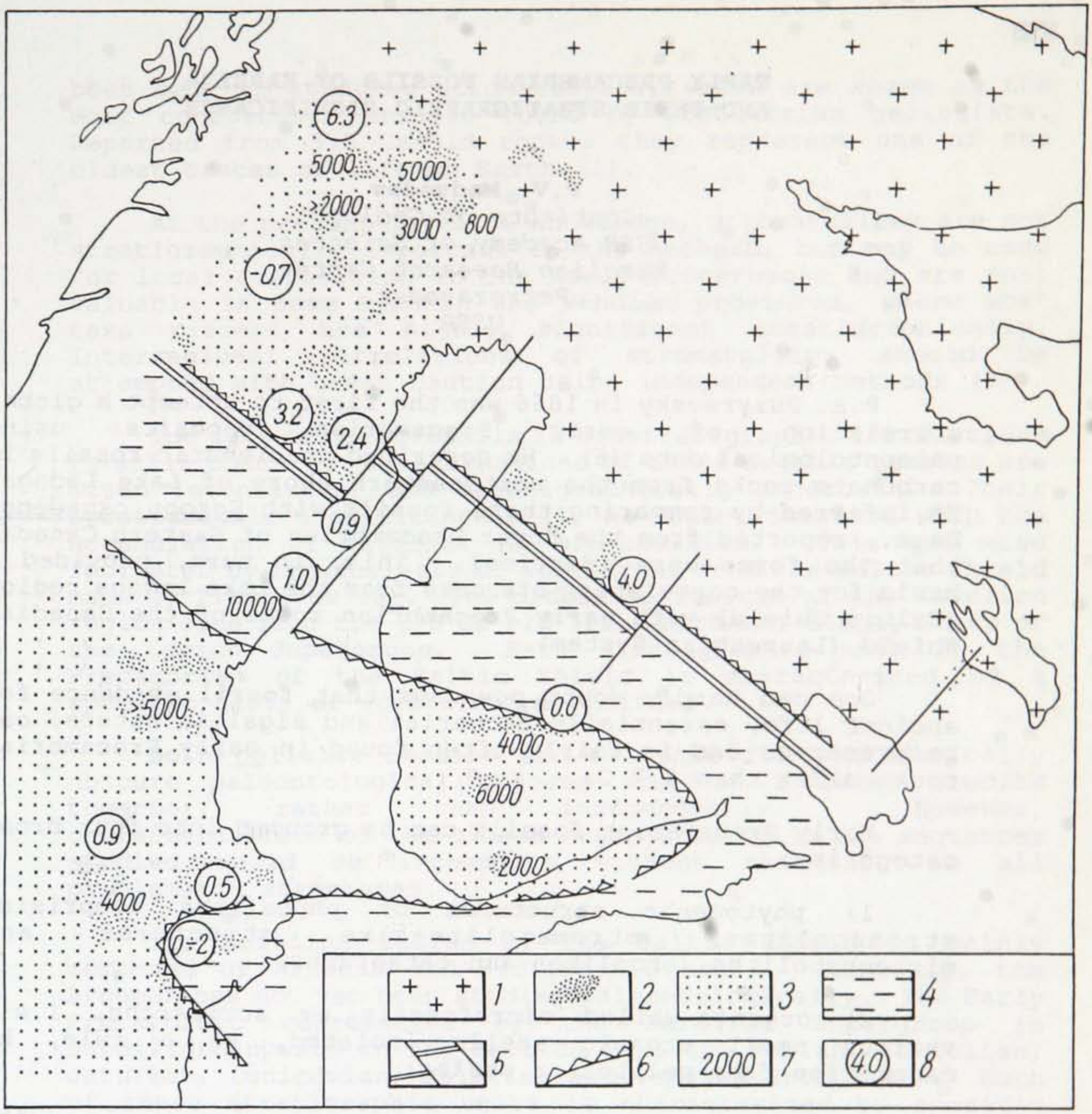


Fig. 4. Svecofennian geodynamic system.

1 - cratonized area of erosion; 2 - investigated part of volcanic arc; 3 - reconstructed volcanic arc; 4 - area of oceanic sedimentation; 5 - spreading axis and transform faults; 6 - subduction zone; 7 - thickness of deposit; 8 - $E_{Nd}$ values.

# EARLY PRECAMBRIAN FOSSILS OF KARELIA AND THEIR STRATIGRAPHIC SIGNIFICANCE

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P.A. Puzyrevsky in 1866 was the first to attempt a global correlation of early Precambrian deposits using paleontological data (6). He described problematic fossils in carbonate rocks from the northeastern shore of Lake Ladoga. He inferred by comparing these fossils with Eozoon canadense Daus., reported from the Lower Precambrian of Eastern Canada, that the forms were identical. This, in turn, provided a basis for the correlation of rocks from the Lake Ladoga region (Baltic Shield) with early Precambrian rocks of the Canadian Shield (Laurentian System).

One can hardly doubt nowadays that fossil evidence for ancient life, essentially bacterial and algal in nature, can be preserved and is fairly often found in early Precambrian rocks older than 1.7 Ga.

Early Precambrian fossils can be grouped into four broad categories:

1) phytogenic structures or phytoliths comprising stromatolites, stromatolite-like structures and microphytoliths (oncolites and catagraphs);

2) organic-walled microfossils or acritarchs, i.e., various small organic shells isolated, as a rule, by maceration from pelites or shales;

3) cellular remains which occur as microscopic units studied in transparent thin sections;

4) problematics represented by systematically obscure biogenic units, including traces and relict organic matter.

Early Precambrian rocks of the Baltic Shield contain all four categories of fossils. Pre-Riphean fossils have so far been reported from about 60 localities in the Karelian region of the Baltic Shield (3).

The Institute of Geology in Petrozavodsk has been conducting systematic paleontological investigations since the late 1950's. Of the above categories of fossils, phytoliths have been studied most thoroughly. A tremendous success has been made in studying stromatolites, which are known as the most conspicuous fossils found by Precambrian geologists. Reported from 3.5 Ga-old rocks, they represent one of the oldest traces of life on Earth (1).

At the present level of knowledge, stromatolites are not stratigraphically important in the Archean, but may be used for local correlation in the Early Proterozoic and are most valuable in some Riphean and Vendian provinces, where most taxa present are highly significant stratigraphically. Interregional correlations of stromatolites should be attempted with great caution using independent methods (7).

The role of microfossils in biostratigraphic correlations is still obscure. Determinations of these organic remains are often inconsistent and stratigraphically uncertain. Their stratigraphic significance will, no doubt, increase with the accumulation of material on microfossils. This was also indicated by the analysis of acritarchs from the Baltic Shield made by Timofeyev (8). The oldest acritarchs of the Karelian region come from the 2.8 Ga-old (late Archean) Gimoly Group of the Lopian Supergroup. Each stratigraphic unit in the Precambrian of the Baltic Shield is characterized by a specific habit of acritarch communities.

Both cellular remains and evidence for systematically obscure paleontological objects are, as a rule, studied together, rather than independently. However, characterization of the oldest organic world in the sequences studied would be incomplete without referring to all problematic structures.

The Karelian region of the Baltic Shield is mainly composed of Archean and Proterozoic rocks. Actually, the Archean has not yet been studied paleontologically. The Early Proterozoic represented by the Karelian Supergroup is subdivided upward in the section into the Sumian, Sariolian, Jatulian, Ludicovian, Kalevian and Vepsian Groups (2). Each of these stratigraphic units is characterized by specific phytolith associations. The greatest number of stromatolites have been reported from carbonate rocks of the 2.45 - 2.25 Ga Jatulian Group.

The detailed study of fossil assemblages provides a basis for subdividing the lithologically uniform Jatulian carbonate sequence into five stratons ranking as beds with phytoliths (from bottom upwards): <u>Nuclephyton</u>, <u>Sundosia</u>, <u>Omachtenia kintsiensis</u>, <u>Butinella</u> and <u>Kalevia ruokanensis</u> (3,4). The beds are given the names of distinctive stromatolites. Two of the above stratons--the <u>Omachtenia kintsiensis</u> and the <u>Butinella</u>-bearing beds--have been traced for a few hundred kilometers in the Karelian region of the Baltic Shield. Stromatolites and stromatolite-like structures are less common in rocks underlying and overlying the Jatulian. They have been encountered in the interval ranging from the Sumian to the Vepsian, inclusive (2.5 - 1.7 Ga).

The distribution of fossils in the early Precambrian rocks of Karelia is fairly uneven. However, most of the above stratigraphic units contain specific characteristic assemblages. This allows the use of paleontological information for the subdivision and correlation of early Precambrian sedimentary sequences (Figs. 1-4). The use of fossils for detailed subdivision is mainly restricted to the Jatulian Group.

The unique pattern of the Precambrian should be taken into account when discussing either the importance of the above categories of fossils for general Early Proterozoic stratigraphy or the role of biostratigraphy at this stage. The stratigraphic units of the Precambrian are not comparable to those of the Phanerozoic, although their taxa are similar in nomenclature. The Precambrian is chronologically remote, and the time boundaries of its units are higher by one order than those of the Phanerozoic. On the other hand, organic life was still at a very low level. The organisms were so primitive that their evolution can only be a matter of speculation. One should also bear in mind the repeated effects of later processes, widespread metamorphism, and other factors which markedly disturb the original pattern of paleocoenoses.

Therefore, the paleontological method employed in stratigraphic studies should be combined with other methods for age determination, and biostratigraphic constructions must be based on all available data on organic remains. It would be premature in this connection to distinguish type complexes and index species. The names of phytolith-bearing beds derived from individual fossils should be regarded as tentative.

Available paleontological information suggests that the evolutionary level of the early Precambrian world was high enough for accumulation of a considerable amount of biogenic matter and for providing a basis for qualitative changes in the subsequent evolution of the biosphere.

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Fig. 1. Problematic organic structures, Jatulian Group, Onegian Formation, lower part, Sundozero Lake region. Thin section view, x6.

Fig. 2. Column stromatolites, Ludikovian Group, Transonegian (Zaonegian) Formation, lower part, Janisjarvi Lake region. Coin is 2.5 cm in diameter.

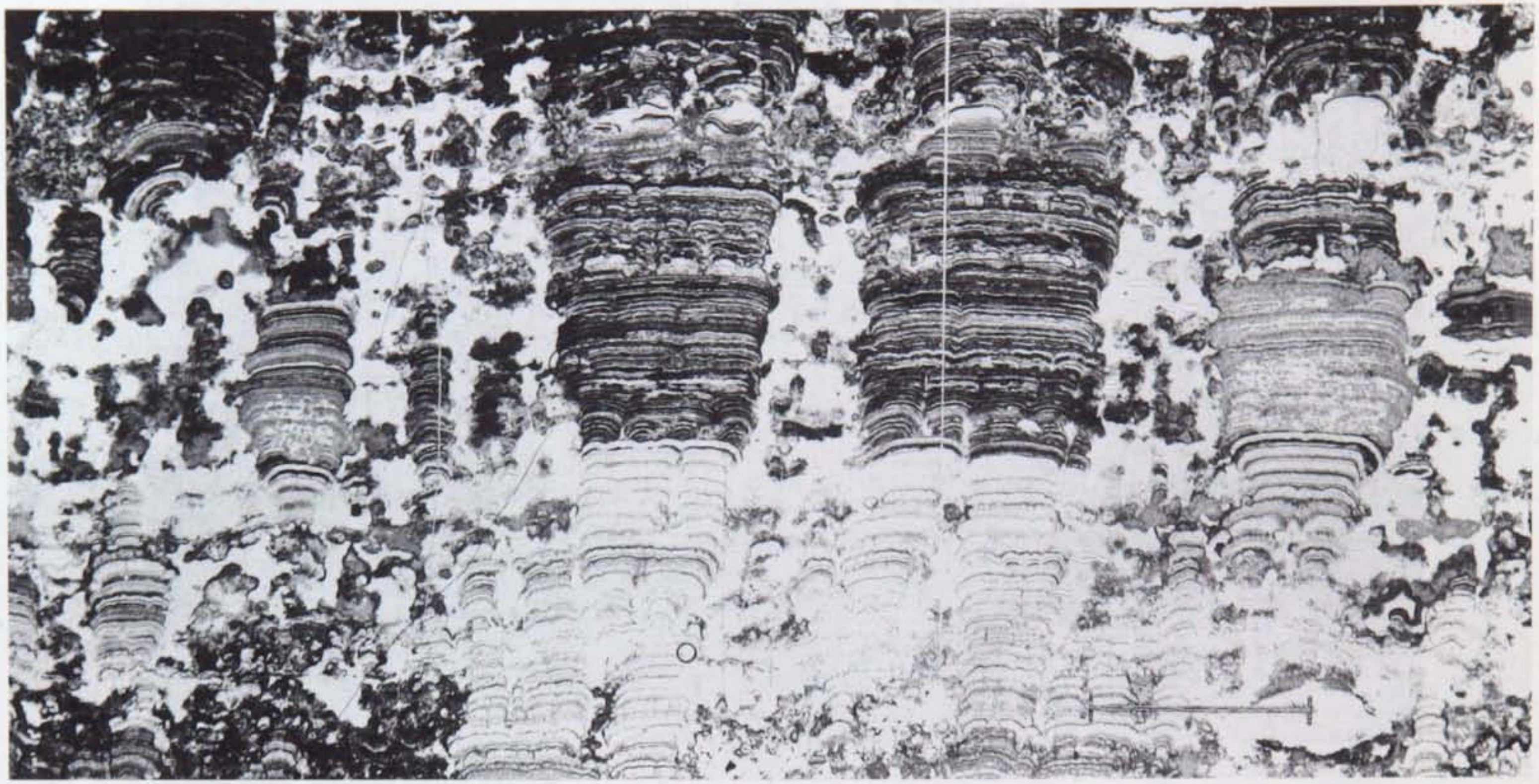

Fig. 3. Primary siliceous microstromatolites (i.e., stizolites), Vespian Group, Vashozero Lake region. Bar is 10 mm long.

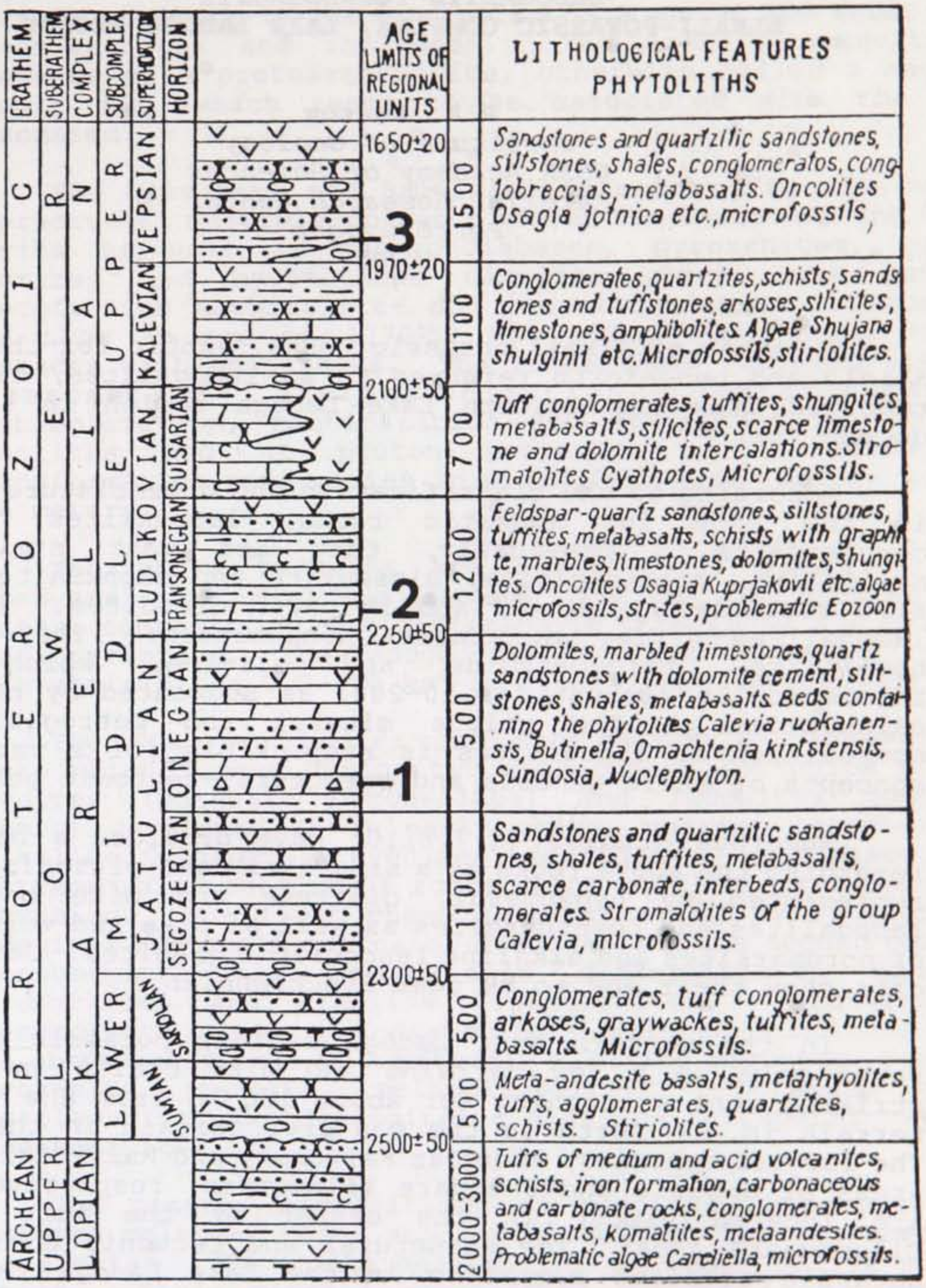


Fig. 4. Stratigraphic column of Karelian Precambrian ($AR_2$ - $PR_1$). 1 = problematic organic structures, Sundozero Lake region (see Fig. 1); 2 = column stromatolites, Janisjarvi Lake region (see Fig. 2); 3 = stizolites (primary siliceous stromatolites) (see Fig. 3).

# LADOGALITE-TOENSBERGITE ALKALI-POTASSIC COMPLEX, LAKE LADOGA REGION

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A number of alkali potassic rocks, exotic for the Baltic Shield and henceforth referred to as ladogalites, have been reported from the western Lake Ladoga region in the past decade (5).

According to the classification and nomenclature adopted in the USSR for magmatic rocks, ladogalites have no counterparts. Chemically, they are most similar to melilitite, melaleucitite, missourite and shonkinite. They also resemble ugandites, and in part, orendites. However, ladogalites differ in mineral composition. They contain nomelilite, feldspathoids and olivine, which could theoretically account for 10-20%, as suggested by normative calculations. The unique mineral and petrogeochemical composition of ladogalites is responsible for a variety of concepts of their genesis and structural-tectonic position.

Numerous studies (5,7,8,10) have provided a basis for regarding the above rocks as a single volcano-plutonic complex represented by hypabyssal, diatreme and dike bodies of ladogalites and toensbergites as well as dike and vein bodies of nordmarkites and alkaline leucocratic granites. Geological data show their age to be post-Svecofennian.

In the western West Ladoga region, potassic alkaline rocks make up a few terrains and dike bodies in a NE - striking zone stretching for about 100 km from the Velimaki terrain in the north to the Ojajarvi terrain in the south. The largest terrains, known as Kaivomaki and Raivimaki, occupy areas of about 8 and 6 square kilometers, respectively, and are located almost in the center of the zone in the Elisenvaara area. The structural and tectonic position of potassic alkaline magmatism in the Lake Ladoga region is determined by the fact that it is confined to the Baltic-White Sea deep-fault belt, which is parallel to the NW margin of the Russian Platform and extends along the SE slope of the Baltic Shield. It has been shown to have been repeatedly active in the Riphean and Paleozoic (3,4,5).

Geophysical studies have revealed a geophysical anomaly in the northwestern Ladoga region. It is characterized by intense heat flow, lowered resistance, a slightly weakened regional magnetic field, increased gravity field values, a certain position of seismic horizons A and M and fault zones (near-vertical and inclined). The anomaly suggests the presence of a protoasthenolite, otherwise called a magmatic diapir (1), which seems to be associated with the rocks discussed.

The Kaivomaki and Raivimaki terrains (Fig. 1) cut the supracrustal multiple-folded rocks of the Sortavala and Ladoga series as well as gabbro-diabases, pyroxenites, gabbro, diorites and hypersthene diorites which have suffered Svecofennian amphibolite- and granulite-facies metamorphism and show an age of $1770 \pm 50$ Ma. They have cross-cutting contacts with host rocks. The latter have suffered alkaline metasomatism (feldspathization, pyroxenization, amphibolization, biotitization and apatitization) both in xenoliths and at pluton contacts. The metasomatic transformation zone varies in width from a few meters to 200-400 meters.

The above terrains are composed of rocks of both hypabyssal and diatreme facies, similar in chemical and mineral compositions but differing in structural and textural characteristics. The ladogalite-toensbergite bodies are cut by breccia pipes 0.2-8.0 $km^2$ in size within the terrains.

The hypabyssal bodies consist of medium-to-coarse-grained rocks which formed during two consecutive intrusion phases. Nevoites (potassic pyroxenites) and ladogites (potassic gabbro) developed at an early phase, whereas toensbergites (alkaline feldspathic syenites) formed at a late phase. The diatremes also consist of rocks which resulted from two-phase intrusion. The first phase is represented by fine-grained ladogites and their explosion breccias and the second phase by toensbergites and their acid differentiates such as nordmarkites and alkaline leucogranites. Ladogite and toensbergite dikes and diatremes developed simultaneously. The above rocks make up a potassic alkaline sequence: potassic pyroxenite (nevoite, melanoladogite) - potassic gabbro (mesoladogite, leucoladogite) - alkaline feldspathic syenite (toensbergite).

The mineral compositions of nevoites, hypabyssal and diatreme melano-, meso- and leucoladogites and toensbergites are similar. The rocks only differ in quantitative mineral ratios. Their mineral composition (in %) is as follows: clinopyroxene (composition: aegirine-augite-salite) 0-65, hornblende and fluorine hastingsite 0-45, composite Ba- Sr- Na K-feldspar (monocline microperthite) 0-80, phlogopite and magnesian biotite 0-45, fluorapatite 0.5-30, albite-oligoclase 0-30, titanite (sphene) 0.2-5.0, orthite 0-5.0 and magnetite 0.2-10.0. Accessories are represented by zircon, xenotime, rutile, ilmenite and olivine.

Hydrothermal minerals such as barite and celestine, associated as a rule with calcite, are seldom found either in individual small grains or in small grain clusters.

The most remarkable components of the rocks discussed are the feldspars (10). They vary quantitatively from 0-10% in nevoites to 80% in toensbergites. Their unique chemical compositions, unusual for native feldspars, seem to explain why they have not been observed in other rocks. As shown by 35 analyses, they contain $K_2O$ (2.87-10.21%), $Na_2O$ (2.25-8.82%), CaO (0.2-1.72%), BaO (0.37-5.80%) and SrO (0.73-5.00%). Other typomorphic minerals of ladogalites such as pyroxenes, micas, amphiboles and apatites are similar in composition to corresponding minerals that make up potassic sequences in other regions (9).

Diatreme ladogites and their explosion breccias constitute breccia pipes up to 850 x 950 $m^2$ in size and oval in plan, and occasional dikes as thick as 30 m. Clinopyroxene, phlogopite, magnesian biotite and apatite megacrystals have been found for the first time in the Baltic Shield in Proterozoic breccia pipes and dikes, together with xenoliths of both host rocks and rocks unknown in the region such as medium- to coarse-grained pyroxenitic rocks and medium- to coarse-grained apatitic rocks that are sometimes monomineralic but more often contain phlogopite, pyroxene or both minerals in varying amounts. The size of the inclusions does not exceed 15-18 cm.

Clinopyroxene megacrystals are most common. Their inclusions vary in size from 10-12 cm to tiny grains (1.0-0.5 mm) indiscernible in fine-grained (diatreme) ladogite matrix. The inclusions are rounded, occasionally angular, with crystal faces seldom preserved. Many crystals contain long-prismatic yellowish apatite inclusions (chadacrysts) or form intergrowths with either phlogopite or magnesian biotite. Pyroxene is present in megacrysts as high-alumina ($Al_2O_3$ up to 9.34%) augite enriched by $Na_20$ (up to 1.8% - jadeite minal), $K_20$ (up to 0.7% - potassic analog of jadeite), $TiO_2$ (up to 0.9%) and SrO (up to 0.18%).

Mica megacrysts are less common, but locally dominate the clastogene constituents of explosion breccia. They occur as dark-brown, almost black, commonly folded and cleaved, irregular, occasionally pseudo-hexagonal crystals up to 6 cm across and ca. 2.5 cm in thickness which contain prismatic apatite crystal inclusions. Some megacrysts are overfilled with acicular rutile. Micas occur in megacrysts as magnesian biotite and occasionally as iron phlogopite, as a rule with increased $TiO_2$ (up to 5%), BaO (up to 3.76%) and F (up to 1.58%) contents. The presence of $P_2O_5$ in some samples is related to fine acicular apatite intergrowths.

The apatite megacrysts are prismatic and light-yellowcolored. They are normally fissured, sometimes crushed and are no longer than 3-4 cm with only one exception (up to 12 cm). Explosion breccia contains both apatite megacrysts proper and apatite chadacrysts in clinopyroxene and mica. It also contains monomineral apatite clasts (inclusions), sometimes with a mica or clinopyroxene admixture. These are fluorapatites with up to 3.3% F, up to 0.27 Cl, up to 1.5% SrO, up to 0.32% $Na_2O$ + $K_2O$ and up to 1.5% REE (all values in % by mass).

Megacrysts, chadacrysts and apatite rock clasts are similar in chemical composition to apatites present in diatreme ladogalites, but the latter have no $K_2O$ and are poorer in $Na_2O$.

Petrochemically, the above rocks are characterized by a low silica content, the predominance of potassium over sodium ($NaO/K_2O$ = 0.35-0.84), and high total alkali (3.5 - 13.2% by mass), $P_2O_5$(0.43-10.5), SrO (0.45-2.2), BaO (0.23-2.0), F (0.12-1.5) and REE (up to 0.5%) contents.

It is seen from Figure 2 that rock-forming oxides are clearly separated in terms of their behaviour in the nevoite-syenite sequence: $SiO_3$, $Al_2O_3$, $Na_2O$ and $K_2O$ contents increase markedly, whereas MgO, FeO, $Fe_2O_3$, $TiO_2$, CaO, $P_2O_5$ and F contents decrease. This agrees with the separation of elements during the silicic-acid differentiation of alkaline basaltic magma.

Isotopic $Sr^{87}/Sr^{86}$ ratios of 0.7038 $\pm$ 0.0005 - 0.7048 $\pm$ 0.0004, the content of K, Na, Rb, Sr, Ba, La, Yb, Ti, Zr, V, Cr, and Ni, their relations in the rocks which make up the terrains, and their distribution in alkaline ultrabasic and alkaline basic rocks, including kimberlites (13), provided a basis for estimating the depth of the magma source at 120-200 km (6) which generated the magma of the above terrains. This is corroborated by both high-pressure megacryst clinopyroxenes from the explosive breccias of ladogalites and available geophysical records (1). The megacrysts observed are characteristic of deep inclusions of either a black clinopyroxenite or the alumina-augite group. Like in the Phanerozoic, their formation seems to be related to the metasomatism of the mantle caused by a KREEP- type substance flowing from its asthenospheric zone. Furthermore, their formation was accompanied by the development of alkali-potassic basaltic melts. In this case megacrysts are regarded as indicators of deep processes of mantle substance transformation. Some writers (2,12) have shown that a crystalline aggregate, which consists of either olivine, leucite, pyroxene and potash feldspar or phlogopite (magnesian biotite), potash feldspar, pyroxene and a small amount of olivine, may be formed in the course of crystallization from the same alkaline melt under different physicochemical conditions. It seems in this connection that the ladogalite-toensbergite series resulted from crystal differentiation of alkali-potassic basaltic magma enriched by REE, Sr, Ba, $H_2O$, $P_2O_5$, HF, HCl and, to a lesser extent, $CO_2$.

Finally, ladogalites and toensbergites are of interest as a new type of magmatic deposit that can be utilized as a source of strontium- and barium-bearing feldspathic, REE-titanium and apatite raw materials. Apatitic, strontium- and barium-bearing feldspathic, REE-titanium and biotite-phlogopite concentrates have been produced in the course of technological laboratory studies.

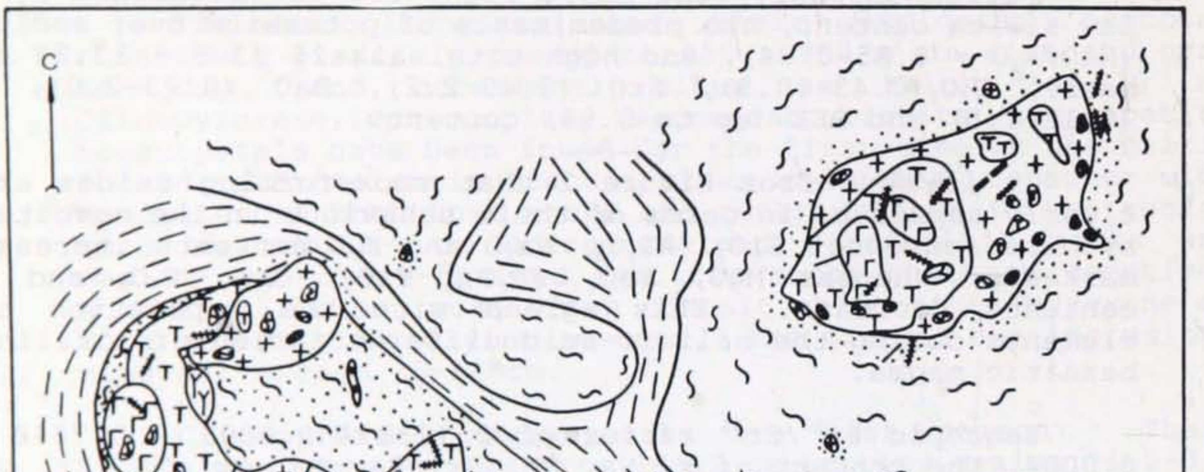


Fig. 1. Geological structure of Raivimaki and Kaivimaki terrains.

1 - syenites, quartz syenite (nordmarkite), syenite-aplite, syenite-pegmatite and alkaline granite veins and dykes; 2 - a) diatreme ladogalites, b) ladogalite dikes; 3 - hypabyssal syenites (toensbergites); 4 - gabbro- fenitized diorites; 6 - quartz-biotite, garnet-biotite-quartz and biotite-quartz schists of the Ladoga series; 7 - amphibole- and graphite-bearing schists and amphibolites of the Sortavala series; 8 - fenitization; 9 - rock boundaries.

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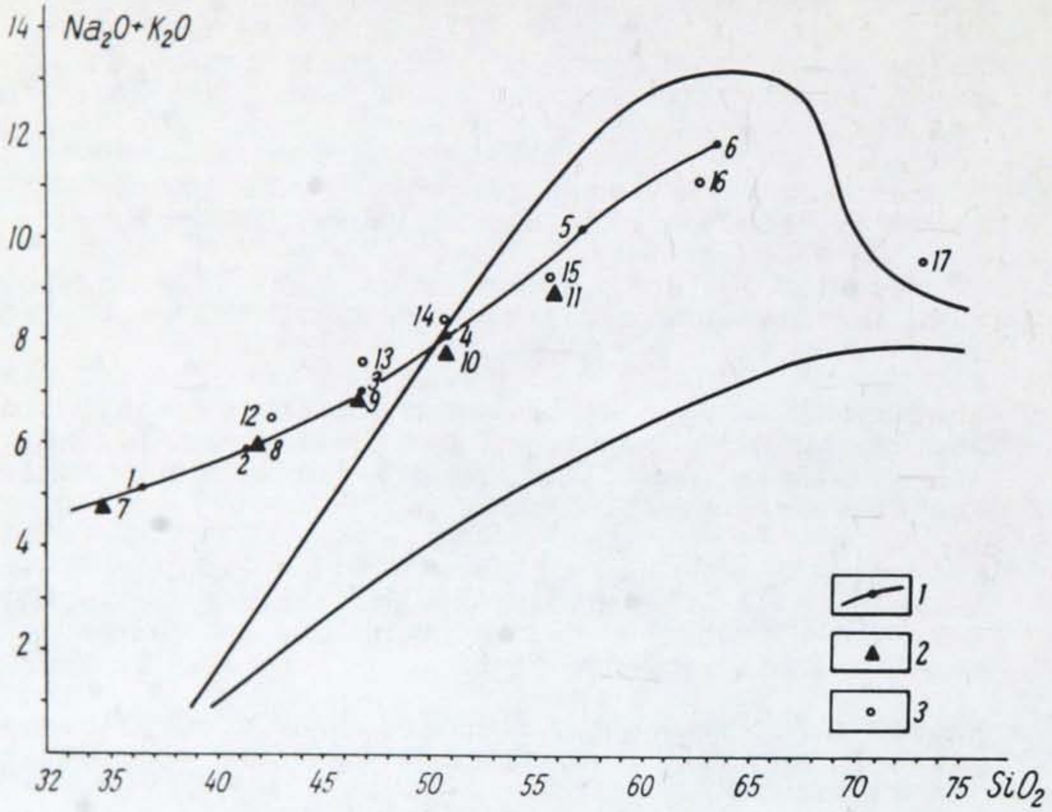


Fig. 2. $Na_2O + K_2O/SiO_2$ ratios in the rocks of the ladogalite-toensbergite complex and their differentiation trend.

1 - hypabyssal rocks and their differentiation; 2 - diatreme rocks; 3 - dike rocks. The numbers indicate: 1,7 - nevoites; 2,8,12 - melanoladogites; 3,9,13 - mesoladogites; 4,10,14 - leucoladogites; 5,11,15 - toensbergites; 6,16 - nordmarkites; 17 - alkaline granite.

# METALLOGENIC SPECIALIZATION OF EARLY PROTEROZOIC VOLCANOGENIC COMPLEXES IN KARELIA

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Copper, nickel, chromium and titanium concentrations have been found in the igneous rocks of different facies that make up the Early Proterozoic complexes. It has been shown by geochemical analyses that copper is confined to both the Jatulian and Zaonezhsky volcanic complexes, nickel and chromium to the Suisarian volcanic complex and titanium to Jatulian subvolcanic rocks. Such geochemical specialization is supported by numerous geological observations. The Zaonezhsky and Jatulian volcanic complexes contain small copper deposits of volcanogenic-phenocryst (bornite-chalcocite-chalcopyrite metallization in lava and subvolcanic strata), volcano-sedimentary (cupriferous sandstones and slates) and hydrothermal (veined, native copper-bearing type). Segregation-magmatic (syngenetic) titanomagnetite vanadium-bearing ore deposits are restricted to Upper Jatulian subvolcanic strata. The Suisarian volcanic complex, notably its subvolcanic rocks, are most promising for copper-nickel and chromite metallization.

Long-term geological studies on Proterozoic volcanogenic rocks of different facies indicate an abundance of localities in which copper and other types of metallization have been observed. Data on copper distribution in both volcanogenic and volcano-sedimentary deposits of different facies have been obtained by studying Jatulian deposits paleovolcanologically.

The study of the mineral composition, morphological characteristics, mode of occurrence and formation of small copper and chalcopyrite deposits and their host rocks such as basalt lava flows and sheets, gabbro-dolerites in sills and dikes, and silicites and quartzitic sandstones in chemogenic and sedimentary-terrigenous members and horizons, provide a basis for dividing copper occurrences and ore deposits into volcanogenic-disseminated, exhalative-sedimentary and hydrothermal types.

Volcanogenic-disseminated type of copper mineralization.

It has been shown by studying copper distribution in the lava fields of Middle Jatulian volcanic zones that copper generally occurs as metallized zones in the first two lava sheets formed during the first and second volcanic stages. The sheets derived from the more basic differentiates of outflowing tholeiitic-basalt lavas contain copper (Table 1).

In Jatulian lava fields, copper concentrations increase in the strata adjacent to the focus of eruption. Away from the focus, the average metal content decreases.

Cuprous sulphides occur as chalcopyrite, bornite and mainly chalcocite, and constitute 1-3% of the rock. These are closely associated with some oxidation-zone minerals such as covellite, cuprite, native copper, malachite and limonite which occur in small amounts.

Ore minerals, e.g., bornite, chalcocite, chalcopyrite and subordinate pyrite, are either disseminated in metallized zones or occur as filler minerals in rounded to isometric amygdules 0.1-0.3 mm in size; amygdules as large as 1 cm in diameter are less common.

Area zoning has been revealed in the paragenetic ore mineral associations of copper in the sequences, far from the assumed eruption center, by studying the Medvezhyegorsk volcanic zone. Essentially chalcocite dissemination is observed in ore basalts which are closest to the center of lava outflow. Bornite mineralization with a small amount of chalcocite is predominant in the ore zones of the more remote sequences. An abundance of disseminated ore is observed along the lava field margins, but it occurs largely as chalcopyrite, chalcopyrite-pyrite and pyrite.

Intraflow ore zoning is also interesting. The epidotized hornblende porphyry basalts, present in the lower parts of the ore zones, show finely dispersed chalcocite dissemination which grades upward into schlieric bornite dissemination that constitutes the bulk of the metallized rock portion. Chalcopyrite grains are often observed in disseminated bornite. The upper parts of ore diabases are composed of amygdaloidal porphyrites, predominantly with both chalcopyrite and pyrite. The total thickness of metallized basalts is 2.5-3.0 m.

The paragenesis of cuprous sulphide minerals in ore zones that coexist with both ferrous/ferric sulphides and ferrous/ferric oxides as well as the areal zoning of metallization suggest that an iron-to-copper ion concentration ratio and partial sulphur and oxygen pressures play a leading role in mineral formation. High-sulphur cuprous sulphide minerals were gradually replaced by lower-sulphur higher-iron types when partial sulphur pressure declined, the Cu/Fe ratio decreased, and oxygen content increased in lava sheets away from the centers of volcanic activity during degassing of outflowing lava.

The migration, concentration and deposition of copper in basic effusive rocks are largely caused by deep magmatic and emanation intraflow differentiation. It is responsible for copper accumulation in the apical parts of magmatic chambers and subsequently in lavas erupted at the initial stages of volcanic activity.

A similar type of copper mineralization has been reported by many workers from basalt terrains of different age, e.g., the basaltic lavas that make up the Umbino suite of the Imandra-Varzuga zone in the Kola Peninsula, the basalts of a trap complex in Siberia, and in the unique copper deposits of Michigan (USA) and Chile.

Cuprous sulphide dissemination also has been found in Jatulian cross-cutting gabbro-dolerite sills.

Chemical analyses of the samples collected in metallized zones with visible chalcocite dissemination have shown that the amount of copper oxide varies from 0.58 to 1.80% (Table 2). The copper content of gabbro-dolerites outside the metallization zones is only a few hundredths of one percent.

## Volcano-sedimentary (exhalative-sedimentary) type of cuprous sulphide mineralization

This type includes native copper and chalcopyrite mineralization in the sandstones, quartzitic sandstones and quartzose gravelstones that constitute the base of the Middle Jatulian unit, where the matrix of terrigenous-clastic rocks contains an abundance of ore minerals such as chalcopyrite, bornite, and chalcocite. The Voronov Bor deposit in the Medvezhyegorsk volcanic zone is most interesting (Table 3). It was formed as a result of post-volcanic fumarole-hydrosolfataric activity at the end of the Early Jatulian phase of volcanism which continued for some time until Middle Jatulian terrigenous-clastic rocks began to accumulate in a shallow-water continental-facies setting. A large quantity of both silica and metals (copper) was removed by mineralized water that contained dissolved gases.

Metallization is confined to the rudaceous quartz gravelstone and sandstone member (0.5 to 1.5 m thick) which rests directly on the rough surface of the underlying Lower Jatulian lava sheet. The top of the lava sheet in the Voronov Bor deposit area is composed of foamy amygdaloidal basalts that form a slaggy, gently undulating surface with depressions occupied by lenticular 0.3-0.4-meter-thick chemogenic volcano-sedimentary jasperoid silicate bodies. The highest degree of ore dissemination in terrigenous rudaceous rocks is restricted to both indistinctly bedded and cross-bedded gravelstones in which copper-bearing ore minerals act as a matrix together with fine clastic and chemogenic quartz material. Dissemination of ore gradually fades out and dissappears upward in fine- to medium-grained quartzitic sandstones and slates.

Chalcopyrite mineralization, similar in genesis, is confined to carbonaceous shales in the Zaonezhsky sedimentary-volcanogenic complex.

The volcanic emanations of Zaonezhsky volcanism are a source of mineralizing fluids in carbonaceous rocks. The hydrothermal field, which formed during the second, most active, phase of Zaonezhsky volcanism, has distinct boundaries on the facies profile among the submarine deposits in the area of active submarine volcanism. In the Zeonezhsky lithotype, this zone is represented by the shungitic rocks, which constitute the second bed of the middle productive horizon, and by overlying shungitic tuffites, most promising for the presence of non-ferrous metals. A relationship between ore-forming and volcanic processes, the sheeted shape of 5 to 7 m-thick ore-bodies, the well-preserved primary structures of rocks and ores, the pattern of sulphide mineralization, and other data suggest that the ore-forming processes in the black shales of the Zaonezhsky suite and those in the similar black shales of the Baltic Shield as a whole belong to the same type.

The 'ore' horizons of carbonaceous shales are of pyritic type. The pyrite ores contain chalcopyrite, sphalerite and pyrrhotite in subordinate quantities. Sulphides occur as beds, layer-by-layer dissemination, nests, concretions, globules, rims surrounding rock and mineral clasts, and veins and veinlets. Metallized slates fall into two genetic types: a primary sedimentary type and an epigenetic-concretionary type.

## Hydrothermal type of cuprous sulphide mineralization

The volcanogenic and sedimentary-volcanogenic complexes of various volcanic zones show widespread quartzose, quartz-calcite, calcitic and quartz-albite veins. The selvages and central portions of the veins contain nests of chalcopyrite, chalcopyrite-pyrite and less commonly, chalcopyrite-bornite (with native copper). Most of the veins are spatially related to basic rocks such as lava flows and lava sheet basalts and gabbro-dolerite sills, predominantly in the Jatulian and the Zaonezhsky volcanic complexes, and are commonly controlled by an extensional-joint system. Veins are largely concentrated in the areas adjacent to either volcanic centers or deep-fault zones which control the position of both volcanic centers and gabbro-dolerite sills. Some veins are a few to 15 meters thick and can be traced for tens of meters along strike.

## Iron-formation occurrences

The volcano-sedimentary (exhalative-sedimentary) type also contains numerous iron-formation and slate occurrences in the volcano-sedimentary complexes that formed in Jatulian and Zaonezhsky time. Hematite-bearing rocks are ubiquitous in the volcano-sedimentary terrains. Distinct hematite-bearing streaks also have been reported from both the Lake Onega (Pyalozero and Tivdia) and Lake Tulomozero areas.

Over 40 iron-formation occurrences have been found in the dolomite suite of the Tulomozero structure. The metallized beds vary in thickness from a few centimeters to 2 m. The quartzites consist of rounded quartz grains as well as chlorite, sericite, hematite and martite.

The volcanogenic rocks also contain jasperoid chemogenic lenses and streaks which consist of fine-grained quartz and finely dispersed hematite; these vary in thickness from a few centimeters to 1.5 m. These chemogenic rocks separate lava flows and indicate long interparoxysmal intervals.

Post-volcanic fumarole-hydrosolfataric processes were a source of iron in both sedimentary sequences and volcanogenic rocks. Hematite-bearing schists and quartzites, similar in genesis, have been reported from the volcano-sedimentary units of the Pechenga complex in the Kola Peninsula.

Titanomagnetite mineralization of volcanic complexes

Titanomagnetite mineralization is widespread in all the volcanic complexes discussed. However, it is most common in the Jatulian complex, and ore is occasionally observed to concentrate in subvolcanic rocks. The Late Jatulian subvolcanic bodies in the Koikary-Svyatnavolok area (Girvas volcanic zone) and those in the Rimskoye area (Pudozhgora titanomagnetite ore deposit described in detail by Kratz, 1959) are of practical interest. It can be seen by analyzing the internal structure of ore-bearing intrusions (Table 4) in inequidistant volcanic zones that they are similar although they occur at different structural levels. The Koikary-Svyatnavolok ore intrusion is in the Upper Jatulian volcano-sedimentary sequence, whereas the Pudozhgora intrusion has been found at the granitoid base.

The above ore intrusions have some common characteristics. They are sheeted, dip gently at 10-25°, the ore horizon is confined to the intrusion sequence, and the rocks and ores are similar in mineral and petrochemical composition.

Available data suggest that the above rock association resulted from liquation (i.e., liquid immissibility) which occurred during the chamber evolution of the initial olivine-tholeiite melt affected by oxidized transmagmatic fluids. As a result of liquation, the initial basaltic melt was split into two fluids. One of them was enriched by iron, titanium, vanadium, cobalt, nickel, and the other by silica, alkalies (sodium), phosphorus, barium, strontium and zirconium.

| | 1. | 2. | 3. | 4. | 5. |
|---|---|---|---|---|---|
| $SiO_2$ | 46.66 | 48.50 | 44.60 | 47.85 | 50.23 |
| $TiO_2$ | 1.80 | 1.39 | 1.70 | 1.19 | 1.07 |
| $Al_2O_3$ | 14.65 | 14.82 | 14.73 | 14.32 | 13.54 |
| $Fe_2O_3$ | 5.06 | 3.95 | 3.70 | 5.48 | 7.50 |
| FeO | 6.91 | 4.57 | 8.87 | 6.50 | 2.71 |
| MnO | 0.81 | 0.17 | 0.20 | 0.18 | 0.19 |
| MgO | 8.51 | 3.98 | 7.79 | 5.51 | 2.93 |
| CaO | 6.50 | 12.55 | 6.00 | 11.68 | 16.23 |
| $Na_2O$ | 3.50 | 2.50 | 3.40 | 0.97 | 0.16 |
| $K_2O$ | 0.05 | 0.25 | not found | 0.10 | not found |
| $H_2O$ | 0.17 | 0.12 | 0.09 | 0.14 | 0.14 |
| Unknown | 3.57 | 4.46 | 7.53 | 3.17 | 4.88 |
| $P_2O_5$ | 0.14 | 0.10 | 0.11 | 0.09 | 0.08 |
| CuO | 2.49 | 2.36 | 0.84 | 2.87 | 0.62 |
| CoO | 0.005 | 0.010 | 0.007 | 0.002 | 0.003 |
| NiO | 0.012 | 0.004 | 0.020 | 0.024 | 0.010 |
| $Cr_2O_3$ | 0.030 | 0.026 | 0.020 | 0.034 | 0.038 |
| $V_2O_5$ | 0.030 | 0.050 | 0.033 | 0.060 | 0.020 |
| S | 0.29 | 0.30 | 0.16 | 0.50 | 0.10 |
| Total | 100.41 | 99.96 | 99.72 | 100.42 | 100.40 |

Table 1. Chemical composition of ore basalts in the Segozero and Medvezhyegorsk volcanic zones.

| Oxides | 1. | 2. | 3. | 4. | 5. | 6. | 7. | 8. |
|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 54.02 | 51.26 | 53.50 | 52.18 | 52.56 | 46.72 | 57.14 | 47.34 |
| $TiO_2$ | 1.68 | 1.88 | 1.68 | 1.56 | 1.80 | 4.30 | 1.80 | 2.64 |
| $Al_2O_3$ | 10.78 | 10.67 | 11.49 | 11.81 | 10.96 | 10.36 | 10.73 | 13.10 |
| $Fe_2O_3$ | 7.89 | 9.47 | 6.93 | 7.53 | 8.65 | 8.22 | 5.72 | 6.32 |
| FeO | 4.60 | 4.80 | 4.74 | 5.10 | 4.54 | 12.35 | 7.04 | 10.48 |
| MnO | 0.07 | 0.08 | 0.09 | 0.07 | 0.08 | 0.18 | 0.09 | 0.22 |
| MgO | 9.26 | 9.66 | 9.35 | 9.66 | 9.66 | 3.98 | 4.20 | 4.68 |
| CaO | 2.52 | 3.24 | 2.80 | 3.22 | 2.80 | 6.18 | 3.92 | 7.84 |
| $Na_2O$ | 2.55 | 2.08 | 2.53 | 2.55 | 2.33 | 3.25 | 5.50 | 3.36 |
| $K_2O$ | 1.20 | 1.37 | 1.10 | 1.00 | 0.98 | 1.21 | 0.33 | 0.88 |
| $P_2O_5$ | 0.22 | 0.22 | 0.02 | 0.22 | 0.22 | 0.37 | 0.62 | 0.26 |
| CuO | 0.83 | 0.60 | 0.70 | 1.13 | 0.66 | 0.58 | 1.80 | 0.60 |
| CoO | 0.009 | 0.012 | 0.008 | 0.011 | 0.009 | 0.007 | 0.004 | 0.007 |
| NiO | 0.005 | 0.005 | 0.005 | 0.005 | 0.005 | 0.004 | 0.004 | 0.020 |
| $P_2O_5$ | 0.040 | 0.040 | 0.030 | 0.034 | 0.028 | 0.060 | 0.010 | 0.080 |
| $Cr_2O_3$ | 0.005 | 0.007 | 0.008 | 0.005 | 0.002 | 0.004 | 0.006 | 0.006 |
| S | 0.10 | 0.10 | 0.07 | 0.17 | 0.05 | 0.02 | 0.15 | 0.03 |
| $H_2O$ | 0.12 | 0.03 | 0.05 | 0.02 | 0.09 | 0.20 | 0.16 | 0.08 |
| LOI | 4.33 | 4.93 | 4.71 | 4.22 | 4.97 | 2.15 | 1.54 | 2.42 |
| Total | 100.13 | 100.36 | 99.95 | 100.32 | 100.35 | 100.11 | 100.54 | 100.32 |

Table 2. Chemical composition of metallized gabbro-dolerite sill, Lake Seletskoye

| Degree of mineral distribution | Ore-bearing minerals: Hypogene | Ore-bearing minerals: Hypergene | Ore-free minerals |
|---|---|---|---|
| Major minerals | Chalcopyrite<br>Bornite<br>Chalcocite<br>Molybdenite | Bornite<br>Chalcocite<br>Covellite<br>Hydrogoethite<br>Malachite<br>Azurite | Quartz<br>Muscovite<br>Chlorite<br>Biotite |
| Minor minerals | Magnetite<br>Ilmenite<br>Pyrite<br>Native silver | Hematite<br>Lepidocrocite<br>Native copper<br>Hydrolepidocrocite | Albite<br>Quartz |
| Admixtures | Tetrahedrite<br>Arsenopyrite<br>Hematite | Cuprite<br>Chrysocolla | Bourite<br>Leucoxene |

Table 3. Mineral composition of ores in the Voronov Bor deposit (after V.P. Bondarev)

| Ore intrusions in the Koikary-Svyatnavolok area (based on the author's data) | Ore intrusions in Pudozhgora (after Kratz, 1959) |
|---|---|
| a) fine-grained to aphanitic gabbro-dolorites up to 3-5 m | a) fine-grained aphanitic diabase up to 5 m thick |
| b) medium-grained poikilophilitic gabbro-dolerites up to 5-10 m thick | b) medium-grained gabbro-diabase up to 35-40 m thick |
| c) ore horizon up to 8 m thick | c) ore gabbro-diabase up to 0.5-0.6 m thick |
| d) medium-to coarse-grained gabbro-dolerites up to 35-40 m thick | d) ore horizon proper, up to 10-16 m thick |
| e) coarse-grained leucocratic gabbro-dolerites up to 20-25 m thick | e) coarse-grained ore gabbro-diabase up to 1-2 m thick |
| f) leucocratic gabbro-dolerites (karyalites) up to 3-5 m thick | f) 100-105 m thick ore-free, medium-grained gabbro-diabase |
| g) fine-grained gabbro-dolerites up to 3-5 m thick | g) micropegmatite (diabase pegmatite)-saturated leucocratic gabbro-diabase 0 to 8.0 m in thickness |

Table 4. Internal structure of titaniferous ore intrusions in south Karelia

# METASOMATISM AND METALLIZATION IN THE TECTONIC ZONE BETWEEN THE WHITE SEA GEOBLOCK AND THE KARELIAN GEOBLOCK

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Svecofennian activity reflects the 2200 - 1800 Ma-old Svecofennian tectono-magmatic cycle which covered southern Finland and the Lake Ladoga region. In the tectonic zone between the White Sea and Karelian geoblocks in Karelia, Svecofennian activity was apparent, both structurally and tectonically, and was accompanied by different facies of metamorphism, metasomatism and specific metallogeny.

The above tectonic zone is a series of synclinorium structures that occur in deep-fault zones and are composed of rocks of different ages. The structures are dominated by Archean greenstone belt volcanics and volcano-sedimentary rocks of andesite-dacite-rhyolite composition. The above Archean (2950-2700 Ma) rocks were metamorphosed under high T- and P - conditions (T = 650-750°, P+9.0-10.5 KBar) in the northern part of the White Sea geoblock tectonic zone, and are represented by greenschist facies in the eastern part. The Proterozoic volcano-sedimentary rocks, which occur in the tectonic zone, make up trough-shaped structures. The early preorogenic metamorphism of the Proterozoic rocks did not exceed the greenschist facies.

Metamorphism of Svecofennian (Middle Proterozoic) age resulted from the increased activity of old roughly EW - and NE - trending deep faults. The metamorphism was concomitant with a few folding deformation stages, most pronounced in northern Karelia, and the development of foliation zones in eastern Karelia. Because metamorphic fluids were highly aggressive, metasomatic processes were widespread at the regressive stages. The processes were accompanied by: (1) recrystallization of earlier Archean (e.g., pyrite, copper-nickel sulphide, magnetite and titanomagnetite) metallization and (2) sulphide, gold, and antimony -arsenic metallization directly related to metasomatism.

In northern Karelia, Svecofennian metamorphism reached a high P and T amphibolite facies (the garnet -kyanite-biotite-muscovite subfacies of the almandine amphibolite facies) and was accompanied by high temperature metasomatism (kyanite-sillimanite subfacies, T = 500-600°, P = 6-8 KBar). Svecofennian metamorphism is characterized by intense acidic leaching followed by redeposition of leached components. The acidic leaching stage is remarkable for the formation of kyanite-quartz-facies metasomatites (e.g., staurolite-kyanite-quartz and garnet-kyanite-quartz types) and lower-temperature (T = 400-500°, P = 5KBar) muscovite-quartz metasomatites. The development of the latter is due to the high mobility of potassium. The mobility of basic components, including iron, is favourable for their subsequent redeposition and the formation of conjugated basic metasomatites such as quartz-garnet, quartz-anorthite, quartz-cummingtonite, quartz-anthophyllite, quartz-gedrite and veinlet-disseminated sulphide ores. The above processes were simultaneous with the recrystallization of pyrite metallization, which existed earlier in the region, and the formation of veinlet-disseminated and massive pyrite and pyrrhotite ores. High-temperature metasomatite zones are dominated by arsenopyrite and antimony-arsenic metallization represented by lollingite, arsenopyrite, tennantite, berthierite, gudmundite, native antimony, antimony-bearing arsenopyrite, and marcasite. According to different authors, the temperature range for the development of antimony-arsenic metallization is 550-200°. The antimony-arsenic metallization in northern Karelia is similar to the Seinajoki deposit in the Svecofennian Belt of Finland. However, no relation between metallization in northern Karelia and granite formation has been found because the tectonic zone of the White Sea geoblock is a high-pressure domain. The presence of gold, which has no commercial value in the region, is associated with the pyrite metallization of the quartz-muscovite metasomatite facies, late chalcopyrite-bearing metamorphogenic quartz veins, and partly with arsenopyrite ores.

In eastern Karelia, metamorphism of Svecofennian age occurred under greenschist facies and seldom epidote-amphibolite and amphibolite-facies conditions. The metamorphism was accompanied by the recrystallization of pyrite. Copper-nickel sulphide and titanomagnetite metallization are directly related to Archean rocks. Svecofennian greenschist metamorphism is characterized by low-temperature metasomatism, widespread at the regressive stage, and the predominance of acidic-leaching-facies metasomatites (beresite, listwanite and sericite-quartzose, fuchsite-quartzose, chloritoid-quartzose and metamorphogenic quartz veins) and associated basic metasomatites such as chloritic, biotitic, epidotitic and essentially carbonate metasomatites. The high chemical mobility of all petrogenic and metalliferous elements is related to low-temperature metasomatism. Metalliferous elements are redeposited at the final stages of metasomatism. Poor sulphide dissemination and the presence of increased quantities of gold in the region are associated with beresite-listwanite- and quartz-sericite-facies metasomatites. Chloritoid-quartz metasomatites are accompanied by disseminated and veinlet pyritic and sphalerite-pyrite-pyrrhotite metallization. When acid-stage metasomatism was superimposed on ultrabasic rocks, chlorite-talc-carbonate slates and listwanites were formed and chalcopyrite-pentlandite-pyrrhotite metallization was recrystallized to form pyrite-millerite metallization represented by high-basaltic pyrite, millerite, polydymite, hersdorphite and ulmannite. In eastern Karelia, antimony-arsenic mineralization represented by arsenopyrite, pyrite, berthierite, hersdorphite, ulmannite, tetrahedrite, burronite and, less frequently, antimony, is associated with beresite-listwanite-facies metasomatites and carbonate-quartz veinlets. In eastern Karelia, antimony-arsenic mineralization was formed within a temperature range of 500-200°. The recrystallization of pyrite ores in eastern Karelia is accompanied by the chalcopyrite, galena, sphalerite, altaite and tellurobismuthite enrichment of pyrrhotite and pyrite-pyrrhotite zones.

Evidence for the Svecofennian metallogenic epoch in the tectonic zone between the White Sea and Karelian geoblocks is provided by both geological and geochemical data and absolute age datings made by the $Pb^{207}/Pb^{206}$ - method for galena and by the K/Ar - method for muscovites and fuchsites from listwanites and beresites. The metasomatic and ore-forming processes have been dated at 2.1-1.8 Ga.

The metallogeny of the above tectonic zone depends, firstly, on pyrite, copper-nickel sulphide and magnetite deposits in Archean greenstone belts and, secondly, on the concentration of Sb, As, Bi, Te, Ag, Au, etc. associated with metamorphogenic-metasomatic processes.

# PRECAMBRIAN NONMETALLICS OF KARELIA: CLASSIFICATION AND GEOTECHNOLOGICAL ASSESSMENT

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Nonmetallics have always played an important role in the technical progress of civilizations. Both rocks and individual minerals are taken into account when classifying nonmetals.

In Karelia, which occupies an area of 172.4 X $10^3$ $km^2$, Precambrian nonmetallics are very important. They include deposits of the minerals muscovite, feldspar, quartz, kyanite, apatite, graphite, talc, and garnet, as well as metamorphic, igneous and sedimentary rocks such as granite, marble, gabbro and gabbro-diabase (natural stone), quartz porphyry, halleflinta, talc-chloritic slate, low- and high-carbon shungite, quartzite, picritic basalt and pyroxenite (products of rock casting), and carbonatites (Fig. 1). Some of the above nonmetals are regarded as multi-purpose raw materials.

Nonmetallics continue to be in great demand. A purely commercial approach to the study of mineral products may result in flagrant errors. The study of nonmetalliferous deposits should be based on both theoretical and practical progress made in stratigraphy, tectonics, petrology, magmatism, mineralogy, geochemistry and geophysics. Furthermore, both geologic-geophysical and technological estimations of available resources are highly significant.

The polycyclic evolution of Precambrian zones is responsible for some characteristics of the geological structure of the region. It is possible to reveal the system dynamics of interaction and evolution by elucidating the developmental pattern of nonmetallics. The nonmetalliferous system is largely controlled by lithostratigraphic, magmatic, structural, metamorphogenic and metasomatic factors.

In Karelia, there are good reasons for distinguishing the Belomorian, Karelian and Ladoga geoblocks as the first-order constituents of the crust of the Baltic Shield (1). This provides a basis for the concept of the block structure of Karelia's lithosphere. Each geoblock shows a unique evolution of metamorphic processes which is indicative of various types of metamorphism. According to (5), Karelia is characterized by the following types of metamorphism:

1. Ladoga: low-pressure andalusite-sillimanite type.

2. West Karelian: moderate-pressure kyanite-sillimanite type.

3. Belomorian-Lapland: high-pressure kyanite-sillimanite type.

The latter type, characterized by high pressure and slightly increased geothermal gradient values, was responsible for the unique regional pattern of mineral formation in the Belomorian geoblock. This implies that metamorphic complexes, most sterile geochemically, are related to relatively unique metamorphism. The pegmatites of the muscovite rock association, which contain among other varieties, a type of muscovite used in TV-sets, are generally recognized as rocks indicative of the economic significance of the Belomorian geoblock. Granulite-facies metamorphism, characterized by low partial water pressure in the fluid and regarded as the earliest metamorphism, has been locally reported from the Belomorian geoblock. Increasing temperature and decreasing pressure subsequently gave an impetus to melting processes. As a result the structures of migmatite terrains were formed on a large scale within a quasi-closed geological system without a substantial addition of major rock-forming components. A general evolutionary pattern of fluid regime is characterized by the increased $H_2O$ and decreased $H_2$ content of the fluid and a decline in both the $H_2/H_2O$ ratio and the reduction coefficient.

Another example is based on the geological and technological study of kyanite ores from the Kichano-Hizovaara zone. Mineral occurrences of the Hizovaara kyanite deposit reported from the link zone of the Belomorian and Karelian geoblocks have suffered the Belomorian - Lapland kyanite - sillimanite high-pressure type of metamorphism. Three types of ores are recognized: metamorphogenic, metamorphogenic-metasomatic, and metasomatic (3).

When constructing flow charts and estimating benefication regimes for kyanite ores, the ore types have been found to be technologically heterogeneous. Variable composition and superimposed secondary processes are responsible for a difference in the physico-mechanical properties of the rocks during crushing. Hizovaara kyanites are less easily beneficiated because of graphitization. The kyanite ores of the Kichanskaya Group are more technologically usable. See Appendix 1.

Complicated flow charts and highly toxic reagents are used to beneficiate feldspathic raw materials produced from pegmatites and to isolate microcline, plagioclase and quartz concentrates. Flow charts are simplified and the quality of concentrates is increased if acid volcanogenic rocks, metamorphosed under moderate-pressure conditions in the Karelian geoblock, are used as a source of feldspathic products. A flow chart, which includes flotation of iron-bearing minerals and micas, has been employed to produce quartz-feldspathic concentrates containing no more than 0.2% $Fe_2O_3$. No separation of feldspars is required because the rocks show either sodic (halleflinta) or potassic (quartz porphyry) compositions when in their native states.

The coarse-grained syenites of the differentiated Yeletozero intrusion were subjected to stadial electromagnetic separation on high-intensity field separators to produce concentrates containing 0.10 - 0.15% iron. The technological value of Yeletozero raw materials is largely due to some petrological characteristics of intrusive terrains, notably higher temperature conditions. See Appendices 2 and 3.

Valuation of Karelia's graphite deposits has shown that ores represented by unaltered biotite gneisses and migmatites with alkaline metasomatites are easily beneficiated. The carbon content of concentrates is 90-95%, with 92-97% of the graphite extracted. Graphite ores from diaphthoretic (i.e., retrograde metamorphism) zones show low technological indices. See Appendix 4.

Based on the stepwise geological evolution of the Precambrian in Karelia, three distinct epochs of apatite formation (Lopian, Karelian, and Riphean) are recognized:

Lopian epoch. It has been shown by studying the Ondozero block that increased apatite concentrations are characteristic of gabbro-pyroxenite and gabbro-diorite magmatic formational types. This type of mineralization is most easily beneficiated.

Karelian epoch. This epoch is represented, for example, by apatite-bearing calcitic carbonatites of the Tiksheozero alkaline terrain. Preliminary data obtained for the above terrain corroborate the fact that relatively simple equilibrium pure mineral systems (apatite-calcite, apatite-magnetite-calcite, apatite-magnetite-phlogopite-calcite) are formed in the course of alkaline magma generation in a generally impoverished geochemical setting. Pure mineral systems substantially facilitate beneficiation. A flotation chart was used to isolate apatite concentrates ($P_2O_5$ content 36.0-38.8%, with 60-70% of the $P_2O_5$ being extracted) from carbonatites. Also, calcitic concentrates were produced. See Appendix 5.

Riphean epoch. In the Ladoga geoblock, this epoch is represented by the alkaline rocks of the Elisenvaara Group which, according to (2), belong to the potassic series. It has been shown that barium-and strontium-bearing feldspar and apatite, biotite, sphene and other types of concentrates can be produced by beneficiation.

The unique evolutionary pattern of the Earth's crust in the Karelian geoblocks is taken into consideration when evaluating nonmetallic deposits. The early Archean granite-migmatite areas of ultrametamorphism, which show increased tectono-metamorphic (metasomatic) protoactivity, are characterized by muscovite, feldspar, quartz and kyanite. Some commercial minerals such as apatite, kyanite, graphite and garnet and commercial rocks such as halleflinta and talc-chloritic slates were formed in Archean granite-greenstone complexes. The Karelian complex contains some commercial rocks, such as low- and high-carbon shungites, quartzites, dolomites and marbles. A number of nonmetallics such as apatite-bearing calcitic carbonates and barium- and strontium-bearing feldspars resulted from intrusive activity, as did building materials including granite, gabbro, gabbro-diabase and charnockite.

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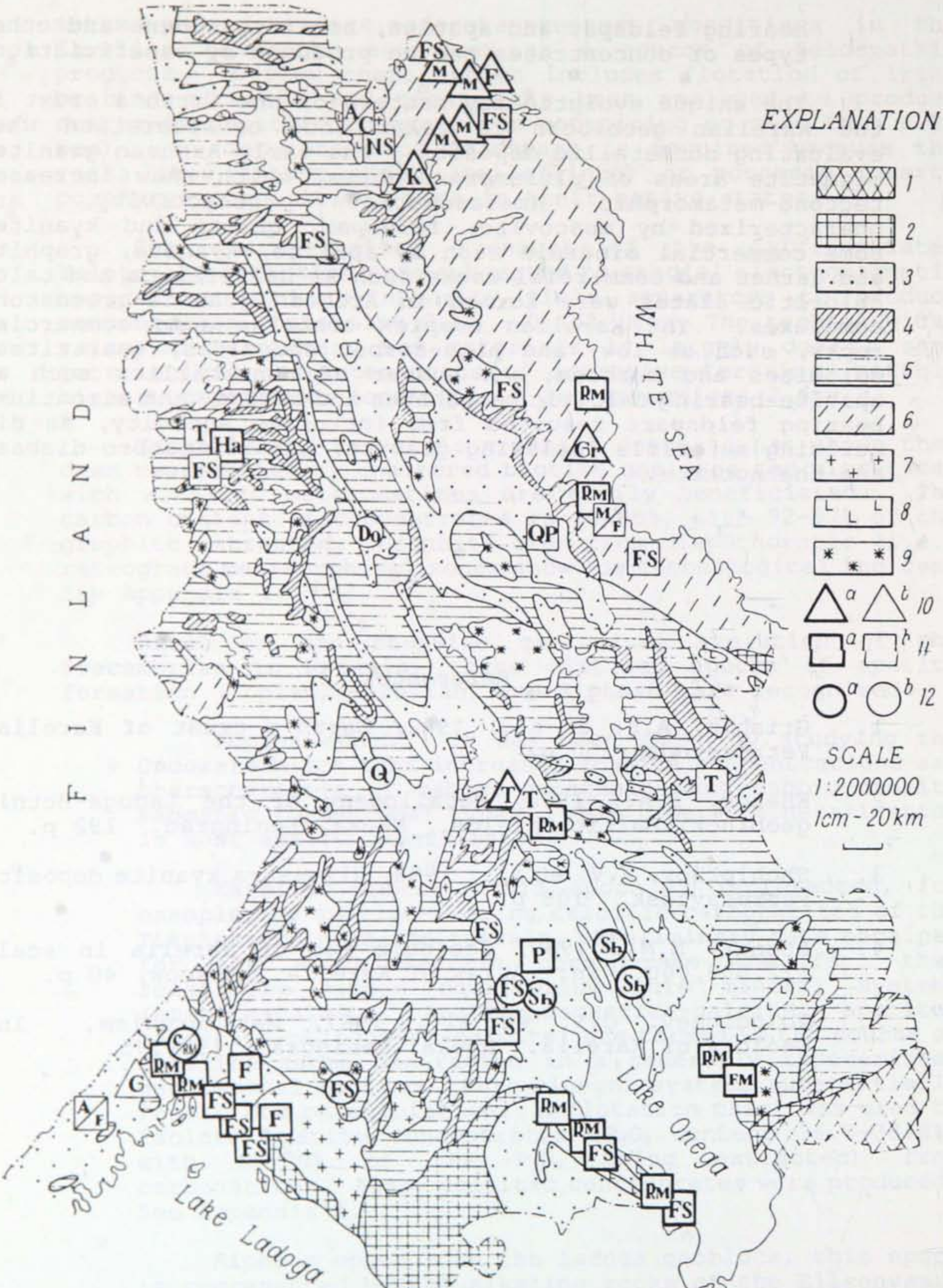
EXPLANATION
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Figure 1. Distribution of major Precambrian nonmetallics in Karelia (Geologic sketch map modified from Stenar, 1989)

1 - Phanerozoic rocks. Precambrian strata: 2 - Vendian and Riphean; 3 - Karelian; 4 - Lopian; 5 - non-divided Lopian and Saamian; 6 - Saamian. Intrusive complexes: 7 - Proterozoic granitoids; 8 - basalts and ultrabasic rocks; 9 - Archean intrusions. Deposits: 10 - metamorphic and ultrametamorphic genesis (a - mined or proven resources, b - potential resources); 11 - magmatic genesis (a - mined or proven resources, b - potential resources); 12 - sedimentary genesis (a - mined or proven resources, b - potential resources). Industrial minerals: K - kyanite, M - muscovite, F - feldspar and quartz-feldspar, Gr - garnet, T - talc, G - graphite, A - apatite, C - calcite. Industrial rocks: Ha - halleflinta, QP - quartz porphyry, NS - nepheline syenite, Q - quartzite, SH - shungite, Do - dolomite, FS - facing stone, RM - road metal, P - pyroxenite.

| Type of metamorphism | Genetic variety | Feature of technological processes | Concentrates, %: Content of $Al_2O_3$ | Content of kyanite | Extraction of kyanite | Example of deposit |
|---|---|---|---|---|---|---|
| kyanite-sillimanite high-pressure | metamorphogenic | flotation | 52-55 | 82-88 | 55-60 | Northern lense |
| kyanite-sillimanite high-pressure | metamorphogenic metasomatic | flotation | 56-57 | 88-90 | 65-70 | Southern lense |
| kyanite-sillimanite high-pressure | metasomatic | magnetic flotation | 53-55<br>52-54 | 84-87<br>82-87 | 60-61<br>58-60 | Eastern and Fuksite |
| kyanite high-pressure | metamorphogenic | magnetic flotation | 57.5-58.5 | 91.5-92.5 | 73-74 | Kichan-skaja group |

Appendix 1. Kyanite ores of Karelia

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 |
|---|---|---|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 57.26 | 57.18 | 57.98 | 59.40 | 59.06 | 59.14 | 60.00 | 57.18 | 56.78 | 59.59 | 59.69 |
| $TiO_2$ | 0.82 | 0.64 | 0.64 | 0.86 | 0.82 | 0.77 | 0.90 | 0.45 | 0.85 | 0.90 | 0.85 |
| $Al_2O_3$ | 19.45 | 19.60 | 19.60 | 16.22 | 16.82 | 16.43 | 18.20 | 20.77 | 19.77 | 18.03 | 17.80 |
| $Fe_2O_3$ | 6.17 | 7.76 | 5.17 | 7.60 | 7.18 | 6.38 | 4.50 | 4.50 | 5.78 | 5.20 | 5.11 |
| CaO | 2.52 | 1.26 | 2.38 | 1.68 | 1.82 | 3.50 | 2.59 | 1.47 | 2.52 | 2.43 | 2.48 |
| MnO | 0.17 | 0.19 | 0.14 | 0.15 | 0.16 | 0.15 | 0.13 | 0.12 | 0.17 | 0.17 | 0.16 |
| MgO | 0.85 | 0.30 | 0.60 | 0.70 | 0.80 | 1.00 | 0.73 | 0.42 | 0.80 | 0.69 | 0.81 |
| $Na_2O$ | 6.45 | 6.17 | 6.45 | 6.17 | 5.71 | 5.77 | 6.31 | 8.15 | 6.66 | 6.08 | 6.15 |
| $K_2O$ | 6.06 | 6.54 | 6.55 | 6.70 | 7.00 | 6.07 | 5.84 | 6.29 | 6.05 | 6.40 | 6.22 |
| $H_2O$ | 0.06 | 0.41 | 0.13 | 0.26 | 0.24 | 0.30 | 0.10 | 0.12 | 0.27 | 0.13 | 0.09 |
| L.o.i. | 0.77 | 0.56 | 0.60 | 0.33 | 0.51 | 0.46 | 0.55 | 0.69 | 0.56 | 0.73 | 0.69 |
| Total | 100.58 | 100.61 | 100.24 | 100.24 | 100.12 | 99.87 | 99.85 | 100.10 | 100.21 | 100.35 | 100.04 |
| $Na_2O+K_2O$ | 12.51 | 12.71 | 13.00 | 12.87 | 12.71 | 11.84 | 12.15 | 14.44 | 12.71 | 12.48 | 12.37 |
| $K_2O:Na_2O$ | 0.94 | 1.06 | 1.02 | 1.09 | 1.23 | 1.05 | 0.92 | 0.77 | 0.91 | 1.05 | 1.01 |

Appendix 2. Chemical composition of alkaline syenite of Eletozero massif (wt. %)

| N | Output | Content | | | | |
|---|---|---|---|---|---|---|
| | | $Fe_2O_3$ | $Na_2O$ | $K_2O$ | $Na_2O+K_2O$ | $K_2O:N_2O$ |
| 1 | 62.28 | 0.25 | 7.40 | 6.71 | 14.11 | 0.91 |
| 2 | 69.80 | 0.30 | 7.40 | 7.46 | 14.86 | 1.01 |
| 3 | 63.51 | 0.27 | 7.43 | 7.32 | 14.75 | 0.99 |
| 4 | 64.61 | 0.20 | 6.94 | 7.04 | 13.98 | 1.01 |
| 5 | 61.40 | 0.27 | 6.57 | 7.42 | 13.99 | 1.13 |
| 6 | 64.40 | 0.26 | 6.94 | 6.67 | 13.61 | 0.96 |
| 7 | 67.40 | 0.23 | 7.40 | 6.53 | 13.93 | 0.88 |
| 8 | 75.10 | 0.34 | 8.94 | 6.53 | 15.47 | 0.73 |
| 9 | 65.90 | 0.20 | 7.00 | 6.31 | 13.31 | 0.90 |
| 10 | 69.70 | 0.16 | 7.00 | 6.90 | 13.90 | 0.99 |
| 11 | 66.80 | 0.10 | 7.00 | 6.67 | 13.67 | 0.95 |

Appendix 3. Technological indices of concentrates from alkaline syenite of Eletozero massif

| | Feature of technological processes | | | Main indices of concentrates, percent | | |
|---|---|---|---|---|---|---|
| Type of metamorphism | Fraction of grinding before the main flotation | Quantity of purifications | Quantity operations of additional grinding | Degree of separation of graphite aggregates | Content of carbon | Extraction of carbon |
| unaltered | 50-60% | 3-5 | 1 | 75-80 | 92.43 | 91.57 |
| Biotite gneisses | class-0.07 mm | | | | | |
| retrograded | " | 5 | 2 | 65-70 | 90.51 | 87.58 |
| Migmatites | " | 3-5 | 1 | 75-78 | 95.00 | 94.58 |
| Alkaline metasomatites | " | 3-5 | 1 | 70-75 | 86.65 | 94.23 |

Appendix 4. Graphite ores of Karelia

| N | $SiO_2$ | $TiO_2$ | $Al_2O_3$ | $Fe_2O_3$ | FeO | MnO | MgO | CaO | $Na_2O$ | $K_2O$ | $H_2O$ | L.o.i. | $P_2O_5$ | S | Total | $CO_2$ |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 1 | 4.87 | 0.39 | 0.61 | 3.05 | 3.12 | 0.20 | 3.55 | 42.85 | 0.47 | 0.39 | 0.18 | 35.15 | 4.90 | 0.19 | 99.63 | 34.0 |
| 2 | 4.56 | 0.31 | 2.19 | 3.51 | 3.23 | 0.093 | 3.28 | 45.36 | 0.49 | 0.38 | 0.12 | 31.02 | 4.48 | 0.02 | 00.62 | 29.84 |
| 3 | 3.62 | 0.16 | 3.18 | 3.75 | 2.72 | 0.215 | 3.78 | 43.76 | 0.31 | 0.57 | 0.19 | 33.21 | 4.32 | 0.17 | 99.78 | 30.34 |

Appendix 5. Average chemical composition of carbonatites of Tiksheozero massif