Age of volcanic rocks and syndepositional iron formations, Marquette Range Supergroup: implications for the tectonic setting of Paleoproterozoic iron formations of the Lake Superior region

D.A. Schneider, M.E. Bickford, W.F. Cannon, K.J. Schulz, and M.A. Hamilton

Abstract: A rhyolite in the Hemlock Formation, a mostly bimodal submarine volcanic deposit that is laterally correlative with the Negaunee Iron-formation, yields a sensitive high-resolution ion microprobe (SHRIMP) U–Pb zircon age of 1874 ± 9 Ma, but also contains inherited Archean zircons as old as 3.8 Ga. This precise age determination for the classic Paleoproterozoic stratigraphic sequence of northern Michigan, the Marquette Range Supergroup (MRS), necessitates modification of previous depositional and tectonic models. Our new data indicate that the Menominee Group, previously ascribed to continental rifting during early, pre-collision phases of the Penokean orogenic cycle, is coeval with arc-related volcanic rocks now preserved as accreted terranes immediately to the south and is more aptly interpreted as a foredeep deposit. We interpret these to be second-order basins created by oblique subduction of the continental margin rather than basins formed on a rifting margin. Along with a recently reported age for the Gunflint Formation in Ontario of 1878 ± 2 Ma, our data suggest that an extensive foredeep in the western Lake Superior region was the locus of iron-formation deposition during arc accretion from the south. Further, we interpret the lower MRS (Chocolay Group), a glaciogenic and shallow-marine succession that lies atop Archean basement, to be equivalent to the upper part of the Huronian Supergroup of Ontario and to represent the original continental rifting and passive-margin phase of the Penokean cycle. The upper MRS (Baraga Group) represents deeper marine basins, dominated by turbidites and lesser volcanic rocks, resulting from increased subsidence and continued collision. A stitching pluton, which cuts correlatives of the Hemlock Formation in a thrust sheet, yielded a U–Pb zircon age of 1833 ± 6 Ma, consistent with other post-tectonic plutons in Wisconsin and northern Michigan, indicating that Penokean convergence lasted no longer than ~40 million years.

Résumé : Une rhyolite de la Formation de Hemlock, un dépôt volcanique sous-marin surtout bimodal qui est latéralement corrélé avec la Formation de fer de Nagaunee, a donné un âge SHRIMP [sensitive high-resolution ion microprobe] U–Pb (zircon) de 1874 ± 9 Ma, mais elle contient des zircons hérités de l'Archéen qui atteignent 3,8 Ga. Cette détermination précise d'âge pour la séquence stratigraphique paléozoïque classique du nord du Michigan, le Supergroupe de Marquette Range (MRS), nécessite donc une modification des modèles de déposition et tectoniques antérieurs. Nos nouvelles données indiquent que le Groupe de Menominee, auparavant assigné à de la dérive continentale au cours de phases pré-collision du cycle orogénique pénokéen, est contemporain avec des roches volcaniques reliées à l'arc, maintenant préservées en tant que terranes accrétés immédiatement au sud; il est plus justement interprété en tant que dépôt d'avant-fosse. Nous les interprétons comme des bassins de second ordre créés par une subduction oblique de la marge continentale plutôt que par des bassins formés sur une marge en dérive. De même qu'un âge récent rapporté pour la Formation de Gunflint en Ontario, soit 1878 ± 2 Ma, nos données suggèrent que la formation de fer se déposait dans une grande avant-fosse dans la région occidentale du lac Supérieur au cours de l'accrétion d'un arc provenant du sud. De plus, nous croyons que le MRS inférieur (Groupe de Chocolay), une succession marine glaciogénique peu profonde qui repose sur le socle archéen, est l'équivalent de la partie supérieure du Supergroupe de l'Huronien en Ontario et qu'il représente la phase originale de dérive continentale et de marge passive du cycle pénokéen. Le MRS supérieur (Groupe de Baraga) représente des bassins marins profonds, dominés par des turbidites et des roches volcaniques moindres, résultant d'un affaissement

Received 16 August 2001. Accepted 25 February 2002. Published on the NRC Research Press Web site at http://cjes.nrc.ca on 17 July 2002.

Paper handled by Associate Editor W.J. Davis.

D.A. Schneider.[1] Department of Geological Sciences, Ohio University, Athens, OH 45701, U.S.A.
M.E. Bickford. Department of Earth Sciences, Syracuse University, Syracuse, NY 13244, U.S.A.
W.F. Cannon and K.J. Schulz. U.S. Geological Survey, MS 954, Reston, VA 20192, U.S.A.
M.A. Hamilton. J. C. Roddick Ion Microprobe (SHRIMP II) Laboratory, Geological Survey of Canada, 601 Booth St., Ottawa, ON K1A 0E8, Canada.

[1]Corresponding author (e-mail: schneidd@ohio.edu).

 DOI: 10.1139/E02-016

accru et de collisions continues. Un pluton de suture, qui recoupe des corrélatifs de la Formation de Hemlock dans une nappe de charriage, a donné un âge U–Pb (zircon) de 1833 ± 6 Ma, ce qui concorde avec d'autres plutons post-tectoniques dans le Wisconsin et le nord du Michigan, indiquant que la convergence pénokéenne n'a pas duré plus que ~40 Ma.

[Traduit par la Rédaction]

Introduction

The Marquette Range Supergroup (MRS) is a Paleoproterozoic continental margin assemblage that lies unconformably upon the southern margin of the Archean Superior craton in northern Michigan and Wisconsin (Fig. 1). The MRS and the broadly correlative Animikie, North Range, and Mille Lacs Groups in Minnesota and western Ontario form the foreland of the Penokean orogen, a ~1300 km long fold-and-thrust belt. The foreland sequences are dominantly epiclastic rocks, with subordinate volcanic rocks and hypabyssal intrusions that are more abundant in the southern part of the foldbelt. The Penokean foreland contains the classic Lake Superior iron ranges (Fig. 1). Studies over the past 150 years have centered on the geology of individual iron ranges and a detailed stratigraphic succession has been developed for each with the aid of abundant data from mines and exploration drilling. The correlation of stratigraphy between ranges has been controversial through to the present day. Incomplete preservation of the Paleoproterozic strata and the widespread blanket of Pleistocene glacial deposits that covers most of the region makes direct physical tracing of units between iron ranges difficult or impossible. A recent summary of the stratigraphy of individual sequences and their correlation is presented in Ojakangas et al. (2001).

A widely held interpretation of the MRS (see Morey 1993 and Ojakangas et al. 2001 for a detailed synthesis) is that it records a progression from intracratonic sedimentation (Chocolay Group) through an Atlantic-style continental rifting and passive-margin stage (Menominee Group) into a foredeep phase (Baraga Group). Foredeep deposition along the southern Superior margin was noted by Hoffman (1987) and Barovich et al. (1989). More recently, it was alternatively proposed that this succession instead developed in a back-arc basin (Hemming et al. 1995; Van Wyck and Johnson 1997). For paleogeographic and geodynamic interpretations (e.g., Karlstrom et al. 2001), it is critical to distinguish foredeep sequences from underlying passive-margin and initial-rift sequences (Hoffman 1987).

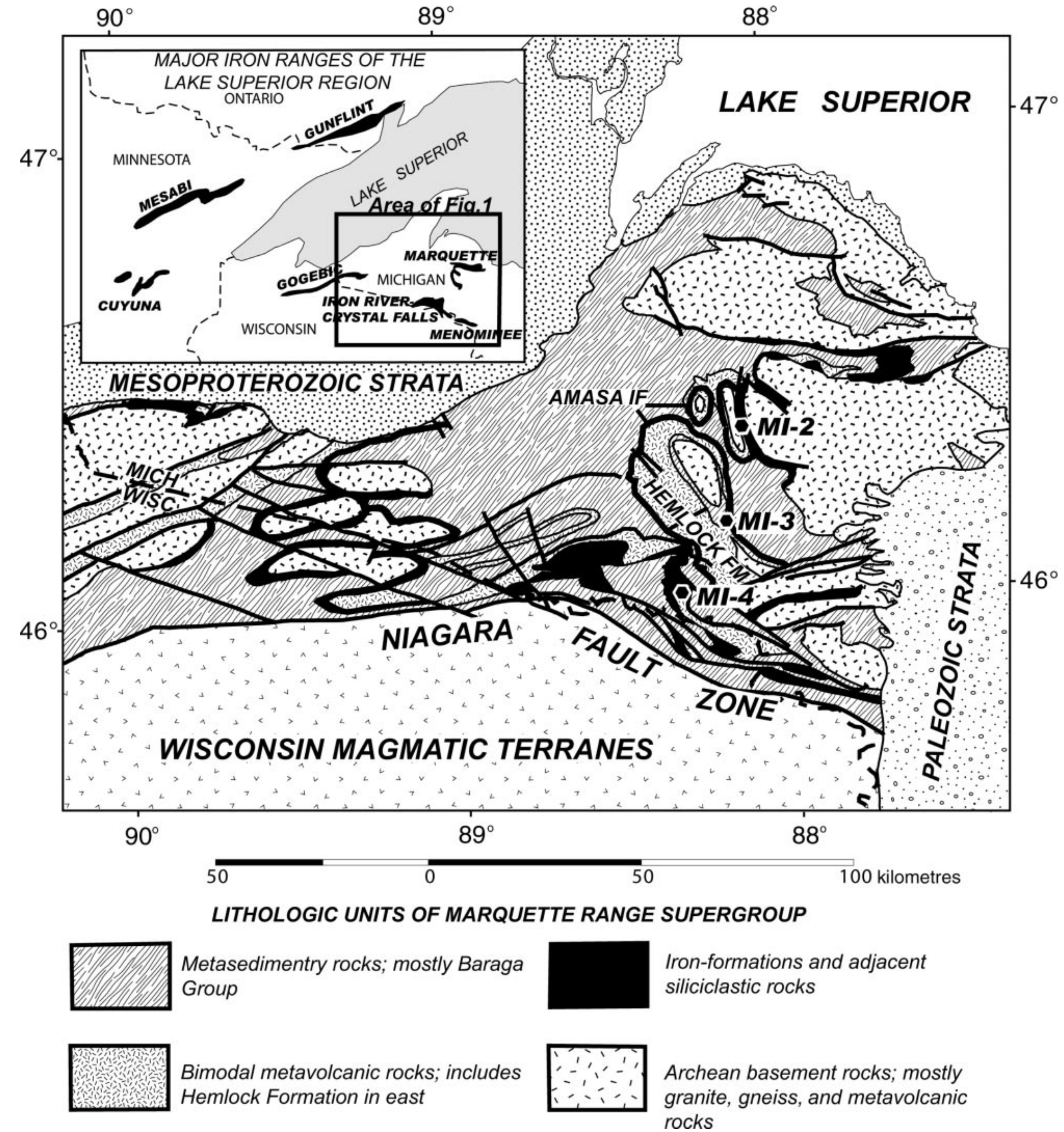
The MRS is bounded on the south by Paleoproterozoic volcano-plutonic, arc-related orogenic rocks of the Wisconsin Magmatic Terranes (WMT), assembled, at least in part, on rifted or dismembered Archean crust (Van Wyck and Johnson 1997). The orogenic system evolved from southward subduction of oceanic crust carrying the Superior plate beneath the microcontinent – island-arc complex. Initial Penokean subduction generated a northerly 1889–1860 Ma volcanic arc (Sims et al. 1992). The oldest volcanic cycles produced tholeiitic basalts and basaltic andesites that are compositionally similar to modern-day island-arc and ophiolitic basalts (Sims et al. 1989). Presumably younger, the calc-alkaline volcanic rocks range in composition from andesite to rhyolite (Sims and Carter 1993). Arc volcanism was accompanied by syntectonic intrusions of tonalite and granodiorite, derived, in part, by partial melting of mantle-derived basalts (Sims et al. 1993).

The Niagara fault (Fig. 1) is generally accepted as the preserved suture between the epicontinental rocks of the MRS and the accreted arc terranes to the south. The Penokean orogeny (1875–1835 Ma; Van Schmus 1976, 1980) is generally attributed to accretion of the Wisconsin arc terrane to the Superior craton. Along this continental margin, the collision produced a foredeep and fold-and-thrust belt that involved Archean basement rocks (e.g., Klasner and Sims 1993) and the formation of gneiss domes (e.g., Schneider et al. 1996).

In marked contrast to the WMT, the tectonic setting and age relations among rocks of the MRS have been less well constrained. Prior to this study, the only published chronostratigraphic limits for the Paleoproterozoic sedimentary units in Michigan and Wisconsin were based on multigrain zircon U–Pb techniques that likely included populations of Archean zircons inherited from subjacent basement. Menominee Group chronostratigraphy was constrained by an age from an interbedded felsic volcanic rock from the Hemlock Formation, which yielded a U–Pb zircon age of 1910 ± 10 Ma (Banks and Van Schmus 1972, cited in Van Schmus and Bickford 1981). Klasner et al. (1989) suggested that sedimentation of the Baraga was underway by 1929 ± 17 Ma on the basis of $^{207}Pb/^{206}Pb$ ages of pebble-size clasts of apatite. A zircon date of 1852 ± 6 Ma (Sims et al. 1989) from a biotite-grade sample of schist from the Baraga Group was interpreted to be the time of cessation of MRS sedimentation, following collision of arc-related rocks of the WMT along the suture zone now marked by the Niagara fault. Chronologic constraints also exist for correlative rocks of the Animikie and North Range Groups in Minnesota, but are based on K–Ar and Rb–Sr mineral ages (e.g., Keighin et al. 1972) that probably record much younger post-orogenic tectonic events. Because of the methodology employed in the older U–Pb analyses (i.e., multigrain zircon populations) and the extent to which K–Ar and Rb–Sr ages could have been overprinted during posttectonic evolution (Holm and Lux 1996), the significance of these ages for constraining the depositional age of the MRS and correlative rocks must be questioned. Although these determinations provided some limits on the age of the MRS, they did not provide adequate precision to correlate stratigraphic sequences with specific tectonic events.

The age reported here for the Hemlock Formation is the first precise U–Pb age for a depositional unit within the MRS and necessitates a reinterpretation of tectonic models for the depositional setting of the supergroup, especially the major Superior-type iron formations and contemporaneous bimodal volcanic rocks. We suggest a new model, in which oblique Penokean subduction formed a complex foredeep north of the Niagara fault zone containing localized extensional features that focussed deposition of volcanic

Fig. 1. General geologic map of the Lake Superior region highlighting Paleoproterozoic Penokean orogenic rocks. Map shows location of samples dated in this study and location of major iron formations of the region (inset). FM, Formation; IF, Iron-formation.

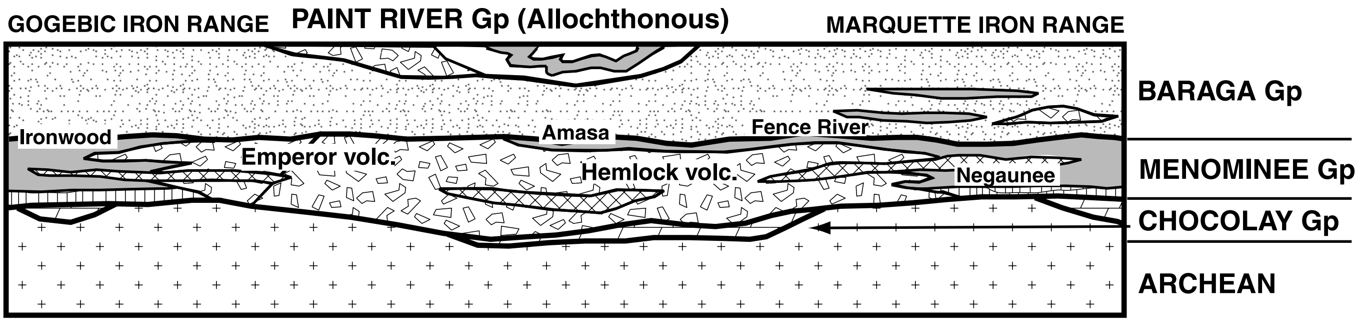


rocks and iron formations. The new date, along with a recently reported age for the Gunflint Formation in Ontario (Fralik et al. 1998), favors the traditional view of contemporaneous deposition of Lake Superior iron formations and verifies that significant magmatism can be generated within continental crust in Paleoproterozoic compressional foredeeps, as originally proposed by Hoffman (1987).

General geology

The Paleoproterozoic supracrustal succession in northern Michigan and Wisconsin is called the Marquette Range Supergroup (Cannon and Gair 1970) and comprises, in ascending stratigraphic order, the Chocolay, Menominee, Baraga, and Paint River groups (Fig. 2). Stratigraphic and sedimentological details can be found in Klasner et al. (1991), Sims and Carter (1993), and Ojakangas et al. (2001). In the following summary, we highlight those characteristics most relevant to later discussions. The oldest group, the Chocolay, is composed of basal glaciogenic sediments overlain conformably by shallow-water arenites, orthoquartzites, and dolomites (Gair 1981). Gair (1981) and Young (1983)

Fig. 2. Diagrammatic pre-deformed stratigraphic relations for the Marquette Range Supergroup from the western Gogebic Range to the eastern Marquette Range, northern Michigan. This figure highlights contemporaneous nature of Ironwood, Amasa, Fence River, and Negaunee iron-formations. Thick black lines represent major unconformities between groups. Patterns: crosses, undifferentiated late Archean; blocks indicate orthoquartzite, carbonate, minor glaciogenic deposits; vertical lines, basal clastic rocks; breccia, volcanic rocks (volc.); gray shading, iron formations; cross-hatching, metagabbro sills; stipple indicate clastic rocks, mostly graywacke and shale. Gp, Group.



suggested that the glaciogenic deposits could be correlated with the Gowganda Formation in Ontario, which lies within the upper part of the 2.4–2.2 Ga Huronian Supergroup. A minimum age for the Huronian is established by the cross-cutting ~2200 Ma Nipissing diabase (Corfu and Andrews 1986). If the correlation for the lower part of the Chocolay with the Gowganda is correct, it would indicate that the Chocolay is likely older than ~2200 Ma.

Lying unconformably on the Chocolay is the Menominee Group which contains some of the largest deposits of iron ore in the region and is well exposed in the eastern Penokean orogen along the margins of Archean basement uplifts (Fig. 1). Menominee sedimentation began with regional accumulation of intercalated quartz arenite and argillite. Along the uplifts, isolated structural basins, possibly grabens (Cambray et al. 1991), dissect the initial Menominee deposits and contain iron-rich strata. Iron-formation sedimentation was accompanied locally by subaqueous eruption of interbedded basaltic and felsic volcanic rocks. The basalts have geochemical signatures similar to modern-day within-plate continental tholeiites (Ueng et al. 1988). Previous workers suggested that the Menominee Group developed in a regional extensional regime (e.g., Schulz et al. 1993).

Unconformably above the Menominee Group is the Baraga Group, which consists of quartzites grading upward into a thick, turbiditic graywacke–shale sequence that includes the several-kilometre-thick Michigamme Formation. A few volcanic and iron-formation members are interlayered within the Baraga Group. Nd isotopic studies indicate that quartzites in the lower part of the Michigamme Formation were derived from an Archean crustal source (Barovich et al. 1989). In contrast, graywacke from the upper part of the Michigamme Formation was derived from the WMT, as the island arc was accreted to the continental margin (Barovich et al. 1989). Correlatives of the Michigamme Formation in Minnesota are the Virginia and Thomson formations, which are also composed of deep-water turbidite-shale units largely derived from Paleoproterozoic source rocks (Hemming et al. 1995).

Sims (1990) suggested that the Paint River Group, formerly considered the youngest stratigraphic unit of the MRS, is an allochthonous sequence thrust over, and possibly the stratigraphic equivalent of, the Baraga Group (Fig. 2). The lower part of the allochthon, the Badwater Greenstone, may be correlative with the Hemlock Formation of the Menominee Group.

Since at least the early work of James (1954), the MRS has been modeled as evolving from a “stable-shelf” environment, represented by the Chocolay and lower Menominee Groups, to a “geosynclinal” assemblage, represented by the Baraga Group. Contending models for the Baraga Group include a foredeep deposit, formed during collision of the Archean craton with volcanic arcs in a south-facing subduction zone (e.g., Hoffman 1987; Barovich et al. 1989) or a back-arc basin deposit formed behind a north-facing subduction zone (Hemming et al. 1995; Van Wyck and Johnson 1997). Regardless of these interpretations, the deposition of continental tholeiitic basalts and coeval rhyolites and iron formations of the Menominee Group within minor fault-bounded troughs indicates that some degree of extension of the Archean basement was occurring during this phase of MRS deposition (Ueng et al. 1988; Schulz et al. 1993).

Sample description and geochemistry

In an effort to constrain the age of the middle units of the Marquette Range Supergroup, and what arguably may represent the “transition from extension to compressional stresses” (Sims and Carter 1993, p. 43), a suite of samples was collected in northern Michigan from both mafic and felsic volcanic units. Zircons were recovered from two felsic volcanic and one granitic rock samples. Geochemical analyses for these samples are presented in Table 1 and Fig. 3.

Two samples from the Hemlock Formation were processed to separate zircons and to obtain major and trace element compositions. MI-99-2 is a foliated, plagioclase phenocryst-rich rhyolite that interfingers with banded iron formation of the Negaunee – Fence River Formation. The sample consists of a fine-grained groundmass composed of quartz, potassium-feldspar, plagioclase, biotite, and muscovite, with accessory zircon and opaques. The rock has a strong foliation, mostly defined by orientation of biotite. Plagioclase also occurs as aggregates up to 3 mm in diameter, whereas quartz occurs as somewhat rounded to ovoid grains up to 2 mm in diameter. We interpret these

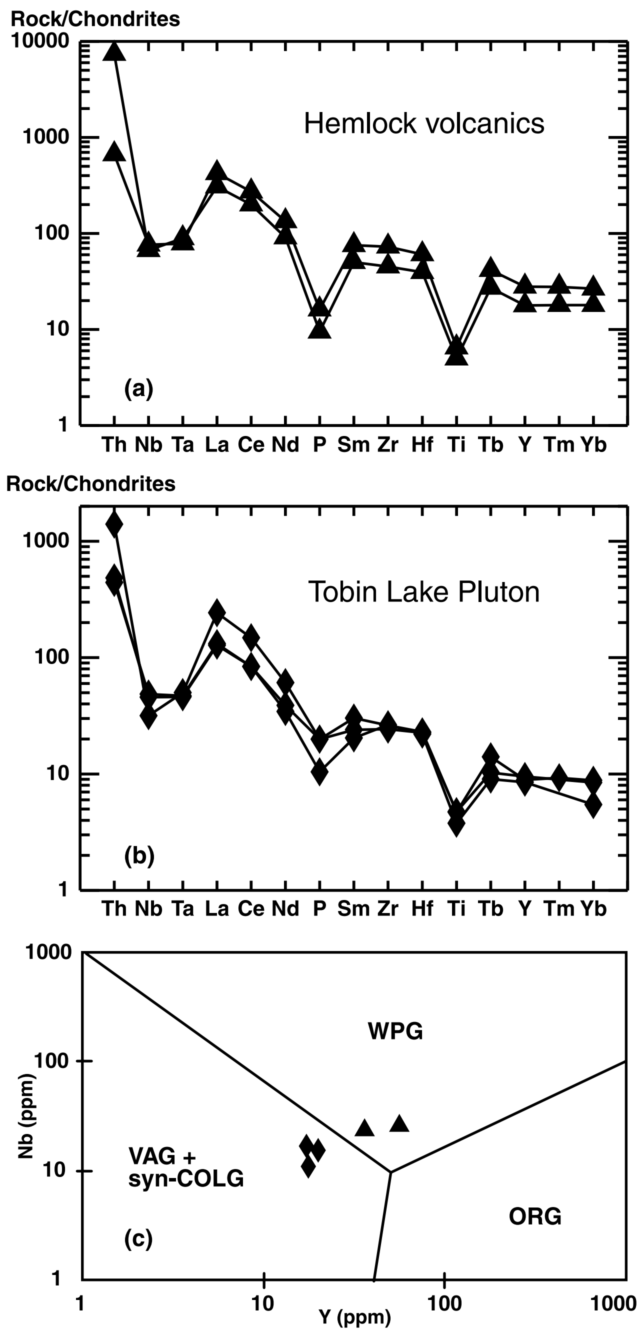
Table 1. Geochemistry of Hemlock rhyolite and Tobin Lake quartz monzonite.

Sample:	MI-99-2	MI-99-3	MI-99-4	802-1	Z515-7A
Location:	Hemlock	Hemlock	Tobin Lake	Tobin Lake	Tobin Lake
SiO_2	73.16	74.86	61.01	61.40	67.40
Al_2O_3	11.63	11.56	15.52	15.40	14.60
Fe_2O_3t	4.11	4.71	5.23	5.84	3.38
MgO	0.73	2.02	3.08	3.14	2.60
CaO	0.83	0.39	1.40	1.67	0.90
Na_2O	2.80	0.14	2.45	2.73	2.99
K_2O	3.88	2.28	6.26	5.55	4.40
TiO_2	0.51	0.67	0.49	0.49	0.39
P_2O_5	0.10	0.17	0.21	0.21	0.11
MnO	0.05	0.03	0.11	0.11	0.04
LOI	1.36	2.86	3.27	3.24	2.00
Total	**99.16**	**99.69**	**99.02**	**99.78**	**98.81**
Sc	10.0	13.0	11.0	10.0	7.0
V	21.0	13.0	68.0	—	—
Co	6.00	6.00	14.0	13.1	10.7
Cr	50.0	28.0	129.0	136.0	85.0
Ni	53.0	—	109.0	33.0	68.0
Rb	88.0	37.0	106.0	99.0	88.0
Sr	57.0	12.0	186.0	217.0	327.0
Cs	0.60	0.30	1.40	1.45	1.50
Ba	1040	190	1010	1058	1130
Y	35.6	55.8	17.8	19.0	17.0
Zr	309	498	179	166.0	181.0
Nb	23.4	26.7	11.1	16.0	17.0
Hf	7.90	12.1	4.60	4.52	4.40
Ta	1.78	1.57	1.00	0.921	0.94
Th	311.0	28.0	58.9	20.3	18.6
U	33.3	2.21	6.31	2.81	2.85
Zn	34.0	47.0	40.0	48.0	52.0
Cu	11.0	—	22.0	20.0	2.0
La	101.0	139.0	80.0	42.0	43.0
Ce	173.0	236.0	128.0	73.0	73.0
Pr	16.3	23.3	11.5	—	—
Nd	57.3	84.3	38.4	24.5	21.7
Sm	10.2	15.3	6.14	4.87	4.14
Eu	1.93	3.53	1.39	0.98	0.83
Gd	8.85	14.2	5.15	—	—
Tb	1.41	2.16	0.73	0.54	0.47
Dy	7.58	12.0	3.77	—	—
Ho	1.50	2.37	0.73	—	—
Er	4.46	6.93	2.14	—	—
Tm	0.61	0.94	0.32	—	—
Yb	3.95	5.85	1.94	1.86	1.20
Lu	0.59	0.83	0.27	0.29	0.18

Note: Major elements listed as wt.%; trace elements listed as ppm. Samples MI-99-2, MI-99-3, and MI-99-4 were analyzed by Activation Laboratories Ltd., Anacaster, Ontario using inductively coupled plasma – mass spectrometry (ICP–MS) and lithium metaborate–tetratborate fusion procedure. Samples 802-1 and Z515-7A were previously analyzed by the U.S. Geological Survey using XRF and INAA techniques (Baedecker 1987).

megacrysts as relict phenocrysts. Other ellipsoidal aggregates of plagioclase, quartz, and micas are up to 5 mm in length and may be relicts of pumice fragments. The second sample is MI-99-3, a blue-quartz-phyric rhyolite. The sample consists of a fine-grained foliated matrix of quartz and feldspar, in which the feldspar component is strongly altered to clay and sericite. Biotite was not observed, but minor opaques occur; zircon and apatite are also accessory minerals. Blue-quartz grains, presumably originally phenocrysts, are euhedral to rounded and 1 to 2 mm in diameter. Elongate

Fig. 3. Extended chondrite-normalized trace element diagrams for (*a*) Hemlock rhyolites and (*b*) Tobin Lake quartz monzonite samples. Chondrite normalization values from Thompson et al. (1983). Note change in scale between Fig. 3*a* and 3*b*. (*c*) Nb–Y discrimination diagrams for granites (after Pearce et al. 1984) showing the fields of volcanic-arc granites (VAG), syn-collisional granites (syn-COLG), within-plate granites (WPG) and ocean-ridge granites (ORG). Triangles represent data from the Hemlock rhyolite samples; diamonds represent data from the Tobin Lake pluton samples.



masses of sericite that are ellipsoidal, up to 1 cm long, and have flame-like ends are probably relicts of flattened pumice fragments. Both samples are characterized by high SiO_2, low Al_2O_3, and relatively high total Fe (Table 1), and are chemically and mineralogically metarhyolites. Variable alkali contents probably reflect secondary mobility during alteration and greenschist-facies metamorphism. Their trace elemental composition are characterized by high Th, U, and light rare-earth element (LREE) contents, and relatively steep negative REE slopes, small negative Eu anomalies (Table 1), and negative Ta, Nb, P, and Ti anomalies on a chondrite- normalized diagram (Fig. 3*a*). The samples plot in the within-plate granite field on the tectonic discrimination diagrams of Pearce et al. (1984) and have the general compositional characteristics of A_2-type felsic magmas (i.e., derived from continental crust or underplated crust) defined by Eby (1992). An origin involving melting of Archean basement is supported by the presence of abundant Archean zircon xenocrysts in the rhyolite samples (see later in the text).

The undeformed Tobin Lake quartz monzonite cuts the strongly deformed rocks of the Paint River Group. Our sample, MI-99-4, is medium-grained (2–3 mm average grain size) and consists of altered plagioclase, microcline, quartz, and somewhat chloritized biotite, with accessory apatite, titanite, and zircon. The rock is hypidiomorphic-granular, with strongly saussuritized, euhedral plagioclase grains intergrown with anhedral microcline and quartz. This sample, and two other analyzed samples of the same pluton are peraluminous (molar $Al_2O_3/(CaO + Na_2O + K_2O)$ = 1.14–1.29) and corundum normative with low SiO_2 and high K_2O contents (Table 1). The trace elements are characterized by relatively high Cr, Ni, Ba, Th, and LREE contents, steep negative REE patterns with only small negative Eu anomalies (Table 1), and negative Ta, Nb, P, and Ti anomalies on a chondrite-normalized diagram (Fig. 3*b*). The sample plots in the field for volcanic-arc-related felsic rocks on the tectonic discrimination diagrams of Pearce et al. (1984). The peraluminous character of the Tobin Lake quartz monzonite and compositional similarity to graywacke–slate units within the Paint River Group (K.J. Schulz, unpublished data) suggest that it was derived by either partial melting and (or) assimilation of sedimentary rocks within the MRS.

U–Pb geochronology

Zircons were separated from the samples by standard methods of crushing, grinding, Wilfley table, and heavy liquid techniques, and analyzed by sensitive high-resolution ion microprobe (SHRIMP II) U–Pb geochronology at the Geological Survey of Canada, Ottawa (GSC). Internal zircon grain structures were imaged by cathodoluminescence and electron backscatter detectors with a Cambridge Instruments S360 SEM (scanning electron microscope), also at the GSC. This technique allowed us to identify older (Archean) xenocrystic cores within the zircons, and either avoid or date them separately from the magmatic populations. Single-grain, spot U–Pb ages were measured following the analytical methods described by Stern (1997). Analyses were carried out with a primary O^- ion beam of ~ 4 nA, focussed to a spot ellipse effecting a sputter diameter of about 13 × 18 μm. Mass resolution was ~ 5100. Ages were determined relative to zircon standard BR266 (559 Ma). Uncertainty in the Pb/U ratios resulting from empirical calibration of the standard was 1.5%. Errors on individual zircon spot ages are reported in Table 2 at the 1σ level; weighted mean ages quoted in the text, however, are presented at the 95% level of confidence (2σ) on the basis of the $^{207}Pb/^{206}Pb$ isotopic ratios.

While both of the rhyolite samples from the Hemlock Formation contained abundant Archean xenocrystic zircon, only MI-99-2 yielded Paleoproterozoic dates (Table 2). Although the grains analyzed were euhedral prismatic crystals, sample MI-99-3 did not yield any Paleoproterozoic dates, from either zircon cores or rims. Rather, zircons from MI-99-3 indicate a spread of $^{207}Pb/^{206}Pb$ ages, at ca. 2640, 2700, and 2810 Ma (Table 2). Because the stratigraphic position of this unit clearly indicates that it is Paleoproterozoic, the zircons analyzed are all interpreted as xenocrysts. Sample MI-2, however, yielded a weighted average $^{207}Pb/^{206}Pb$ age for 8 spots on 6 euhedral, 200 × 50 micron, brown zircons = 1874 ± 9 Ma (MSWD (mean square of weighted deviates) = 0.87; Fig. 4A). This rhyolite also contained Archean xenocrysts with ages ranging from about 2600–2800 Ma; one grain, however, yielded an array of discordant age data whose upper intercept is 3849 ± 75 Ma (MSWD = 0.18; Table 2, but not shown in Fig. 4A). These zircons were presumably inherited from Archean crustal sources, from which rhyolitic magmas were derived or with which they interacted during ascent. The extreme age of 3.8 Ga suggests derivation from rocks correlative with rocks of the Minnesota River Valley terrane (Goldich and Hedge 1974) or the Watersmeet dome (Peterman et al. 1986; Kinny et al. 1991).

The undeformed Tobin Lake quartz monzonite yielded a weighted average $^{207}Pb/^{206}Pb$ age for 9 spots on 9 zircon grains of 1833 ± 6 Ma (MSWD = 1.5; Fig. 4B), placing a younger age constraint on the deposition of the Marquette Range Supergroup, as well as cessation of Penokean-related collision. The absence of Archean xenocrysts and the composition of similar alkali-granites within the Penokean suggest derivation from juvenile sources (Van Wyck and Johnson 1997).

Discussion

Iron-formation correlation

The western Lake Superior region has long been known (and is the type region) for Superior-type iron formations — extensive cherty iron deposits interbedded with mostly clastic sedimentary rocks deposited on continental basement. The region contains seven principal iron ranges (Fig. 1, inset), as well as additional lesser iron-formation units, particularly in northern Michigan and immediately adjacent parts of Wisconsin. Traditionally (e.g., Van Hise and Leith 1911), the major iron formations were considered to represent a widespread, coeval interval of iron deposition. Since the late 1980s, various alternative correlations have been proposed, in which individual ranges were interpreted to have unique and not entirely correlative stratigraphic sequences and hence somewhat diverse ages (e.g., Morey and Van Schmus 1988; Morey and Southwick 1991; Ojakangas et al. 2001). However, the data presented here, and a recently published age for the Gunflint Formation in western Ontario, suggest that lateral equivalence of the major iron formations is still a viable interpretation and, in fact, is consistent with the best available geochronologic data. Fralick et al. (1998) reported a preliminary concordant U–Pb zircon age of 1878 ± 2 Ma for a lapilli tuff interbedded with siliciclastic rocks in the Gunflint Formation, which have been correlated with the Hemlock and Emperor volcanics in Michigan. Our age of 1874 ± 9 Ma for the Hemlock Formation is analytically indistinguishable from the Gunflint age, and shows that iron-bearing sequences at opposite ends of the Penokean foreland are essentially coeval within analytical uncertainty.

The stratigraphic position of the Hemlock Formation is central to our interpretation that the 1874 Ma age dates the deposition of the major iron formations in northern Michigan. The Hemlock Formation and overlying Amasa Formation, a banded iron formation, were first described and named by Clements and Smyth (1899) and were correlated with the Negaunee Iron-formation of the Marquette Iron Range. Much later, the correlation was changed and the Hemlock and Amasa Formations were placed in the slightly younger Baraga Group on the basis of interpretation of local lithologic details (Gair and Wier 1956). That revised correlation was modified by Cannon (1986) who returned to the original interpretation that the Hemlock and Amasa formations are correlative with the Negaunee Iron-formation and other major iron formations of northern Michigan. This interpretation is consistent with regional relationships, especially in the eastern Gogebic Iron Range, where a clear lateral equivalence of the principal iron formation of the range, the Ironwood Iron-formation, and an extensive volcanic unit, the Emperor Volcanic Complex, have been established by detailed mapping (Trent 1973; Klasner et al. 1998). Relationships in these areas are especially significant in demonstrating that Lake Superior-type iron formations can pass laterally into volcanic rocks over distances of only a few kilometres.

The lateral equivalence of the Hemlock–Amasa units and the Negaunee Iron-formation cannot be demonstrated as unequivocally as that of the Ironwood and Emperor units because of intervening structures and poorly exposed areas. The correlation, however, is well grounded in the spatial distribution of units and their relationship to regional unconformities. In the Marquette Iron Range, the Negaunee Iron-formation is devoid of volcanic rocks, although mafic sills are extensive. The western limit of directly traceable Negaunee Iron-formation is the tight synclinal structure, from which we have derived our age determination (Cannon and Klasner 1976). On the east limb of that structure, a typical Marquette Range stratigraphic section, in which the Negaunee Iron-formation lies on a quartzite–argillite sequence that in turn lies on Archean basement, is well documented by exploration drilling and limited bedrock exposures. On the western limb, the Hemlock Formation lies directly on Archean basement and is overlain by a thin iron formation, known as the Fence River Formation. Although Cannon and Klasner (1976) originally inferred the Hemlock Formation to be unconformably above the Negaunee, we now interpret that within this syncline a rapid lateral transition occurs, analogous to the eastern Gogebic Range, in which the major iron-bearing unit (Negaunee) passes laterally into a volcanic sequence with lesser iron formation (Hemlock). This interpretation is consistent with the revised regional correlation of Cannon (1986). Farther west, the Hemlock Formation attains much greater thickness and the interbedded iron formation, as well as the superjacent iron formation, the Amasa, are the lateral equivalent of both the Negaunee and Fence River iron-formations (Cannon 1986).

In the Marquette Iron Range, the Negaunee Iron-formation, and other metasedimentary rocks of the

Table 2. U–Pb isotopic data, Marquette Range Supergroup units.

Spot[a]	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	^{204}Pb (ppb)	f^{206}c (%)	$\frac{^{206}Pb}{^{238}U}$	±1σ	$\frac{^{207}Pb}{^{235}U}$	±1σ
MI-99-2 Hemlock volcanic										
2-8.1	967	70	0.07	825	21	0.06	0.7157	0.0181	36.006	1.951
2-8.6	445	213	0.49	320	7	0.06	0.5480	0.0229	26.952	1.920
2-8.2	276	157	0.59	236	11	0.13	0.6599	0.0119	31.038	1.147
2-8.3	1170	99	0.09	655	17	0.06	0.5054	0.0087	18.493	0.375
2-8.5	783	86	0.11	381	5	0.03	0.4606	0.0070	12.715	0.204
2-16.1	438	134	0.31	236	2	0.02	0.4928	0.0077	12.354	0.203
2-23.1	566	241	0.44	286	2	0.02	0.4507	0.0069	10.806	0.205
2-18.1	122	109	0.92	49	0	0.02	0.3390	0.0055	5.555	0.123
2-15.2	739	238	0.33	243	1	0.01	0.3161	0.0050	5.054	0.094
2-2.2	160	188	1.21	66	0	0.01	0.3236	0.0056	5.154	0.103
2-15.1	369	202	0.56	136	3	0.04	0.3339	0.0052	5.312	0.089
2-6.1	127	119	0.96	51	1	0.05	0.3312	0.0052	5.250	0.096
2-10.2	206	211	1.06	82	3	0.10	0.3244	0.0051	5.129	0.093
2-7.1	133	142	1.11	54	2	0.11	0.3280	0.0053	5.178	0.099
2-6.2	119	95	0.82	46	4	0.21	0.3275	0.0055	5.146	0.121
2-2.1	218	267	1.26	93	1	0.02	0.3327	0.0055	5.219	0.101
2-10.1	197	210	1.10	82	4	0.11	0.3385	0.0053	5.306	0.097
2-3.1	161	173	1.11	65	1	0.05	0.3263	0.0052	5.094	0.100
MI-99-3 Hemlock volcanic										
3-8.1	184	184	1.04	105	4	0.10	0.4407	0.0069	12.175	0.210
3-6.1	142	59	0.43	87	1	0.02	0.5417	0.0099	14.790	0.303
3-22.1	92	107	1.20	65	3	0.11	0.5342	0.0087	13.649	0.282
3-15.1	2031	213	0.11	1094	6	0.01	0.5164	0.0081	12.687	0.213
MI-99-4 Tobin Lake quartz monzonite										
4-3.2	1022	1029	1.04	387	1	0.01	0.3107	0.0052	4.860	0.088
4-6.1	894	965	1.12	358	3	0.02	0.3237	0.0051	5.029	0.088
4-22.1	937	921	1.02	377	3	0.02	0.3304	0.0050	5.119	0.079
4-4.1	1486	1810	1.26	632	4	0.02	0.3334	0.0054	5.161	0.090
4-16.1	890	846	0.98	338	2	0.02	0.3154	0.0049	4.879	0.083
4-2.1	1135	1232	1.12	456	0	0.00	0.3247	0.0049	5.017	0.087
4-3.1	953	930	1.01	383	3	0.02	0.3316	0.0051	5.120	0.083
4-23.1	1300	1792	1.42	564	1	0.01	0.3310	0.0079	5.105	0.172
4-7.1	1715	2667	1.61	767	2	0.01	0.3283	0.0050	5.047	0.079
4-8.1	636	555	0.90	243	6	0.06	0.3229	0.0051	4.949	0.083

Note: Uncertainties reported at 1σ (absolute) and are calculated by numerical propagation of all known sources of error, and data corrected according to procedures outlined in Stern (1997). * Indicates radiogenic Pb; f^{206}c is percent ^{206}Pb contribution from common Pb.

[a]2, sample number; 2-8.1, refers to grain 8, spot 1;.2, etc., refer to a subsequent spot analyses on the same grain.

[b]Conc., concordance = 100 × (^{206}Pb/^{238}U age)/(^{207}Pb/^{206}Pb age).

Menominee Group lie unconformably on dolomite and quartzite of the Chocolay Group and are overlain unconformably by conglomerate and graywacke of the Baraga Group. In the area of our study, the Hemlock and Amasa Formations occupy the same stratigraphic position between these unconformities. The Hemlock Formation lies either on dolomite of the Chocolay Group or directly on Archean basement gneiss. Although exposed only in now-abandoned mines, an unconformity between the Amasa Formation and overlying Baraga Group has been documented (Leith et al. 1935). Thus, the age equivalence of the Hemlock Formation and the Negaunee Iron-formation seems highly likely, and we propose that the age for the Hemlock Formation reported here is also the age of the Negaunee Iron-formation.

Although our data confirm that two geographically separated iron formations of the Lake Superior region are essentially contemporaneous, the correlation of these units with other iron formations of the region must remain interpretive pending acquisition of precise age determinations from each iron range. Because each iron range displays some variations in lithostratigraphy, and because they are physically isolated from each other, a direct tracing of stratigraphic continuity of iron-bearing sequences is not possible and a variety of interpretations is permissible.

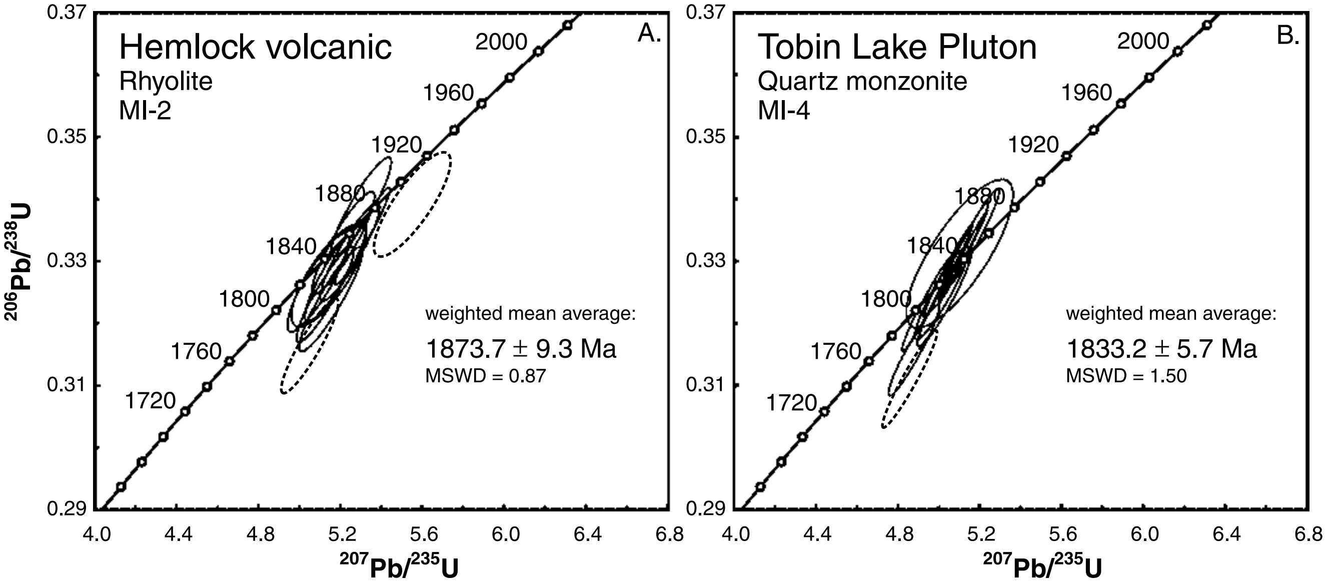
Stratigraphic relationships within 14 different iron-bearing regions, and proposed correlations between regions, was recently presented by Ojakangas et al. (2001). Perhaps the most unifying feature of the somewhat diverse stratigraphic

		Apparent ages (Ma)						
$^{207}Pb/^{206}Pb$	±1σ	$^{206}Pb/^{238}U$	±1σ	$^{207}Pb/^{235}U$	±1σ	$^{207}Pb/^{206}Pb$	±1σ	Conc.[b] (%)
0.3649	0.0164	3480	68	3667	55	3770	70	92.3
0.3567	0.0188	2817	96	3382	72	3736	83	75.4
0.3412	0.0103	3267	46	3520	37	3668	47	89.1
0.2654	0.0023	2637	38	3016	20	3279	14	80.4
0.2002	0.0008	2442	31	2659	15	2828	6	86.4
0.1818	0.0007	2583	33	2632	16	2669	6	96.8
0.1739	0.0016	2398	31	2507	18	2595	16	92.4
0.1189	0.0016	1882	27	1909	19	1939	24	97.0
0.1160	0.0009	1771	25	1828	16	1895	14	93.4
0.1155	0.0009	1807	27	1845	17	1888	15	95.7
0.1154	0.0005	1857	25	1871	14	1886	8	98.5
0.1150	0.0009	1844	25	1861	16	1879	14	98.1
0.1147	0.0008	1811	25	1841	15	1875	13	96.6
0.1145	0.0010	1829	26	1849	16	1872	16	97.7
0.1140	0.0017	1826	27	1844	20	1863	26	98.0
0.1138	0.0010	1851	26	1856	17	1860	15	99.5
0.1137	0.0009	1879	26	1870	16	1859	14	101.1
0.1132	0.0011	1820	26	1835	17	1852	17	98.3
0.2004	0.0011	2354	31	2618	16	2829	9	83.2
0.1980	0.0014	2791	42	2802	20	2810	12	99.3
0.1853	0.0020	2759	37	2726	20	2701	18	102.2
0.1782	0.0008	2684	34	2657	16	2636	7	101.8
0.1135	0.0006	1744	25	1795	15	1855	10	94.0
0.1127	0.0007	1808	25	1824	15	1843	11	98.1
0.1124	0.0002	1840	24	1839	13	1838	3	100.1
0.1123	0.0006	1855	26	1846	15	1837	9	101.0
0.1122	0.0006	1767	24	1799	14	1835	10	96.3
0.1121	0.0007	1812	24	1822	15	1833	12	98.9
0.1120	0.0004	1846	25	1840	14	1832	6	100.8
0.1119	0.0023	1843	38	1837	29	1830	38	100.7
0.1115	0.0003	1830	24	1827	13	1824	5	100.4
0.1112	0.0005	1804	25	1811	14	1818	8	99.2

sequences is the widespread occurrence of a thick sequence of turbidites forming the youngest stratigraphic units. These turbidites, variously called the Michigamme, Copps, and Tyler formations in Michigan and Wisconsin, and the Virginia, Thomson, and Rove formations in Minnesota and Ontario, are widely accepted as deposits of a compressional foreland basin. These units all lie with conformable, disconformable, or low-angle unconformable relationships on the major iron-bearing sequences, indicating that the iron-bearing sequences are only slightly older than the onset of turbidite deposition. This relationship argues for the near contemporaneity of the iron formations, although some uncertainty and disagreement still exists on precise correlations between individual areas.

The correlation of the Gunflint with the Biwabik Iron-formation of the Mesabi Range in Minnesota has long been accepted because the two ranges are essentially along strike and separated from each other by only about 70 km, where Mesoproterozoic intrusive rocks of the Duluth complex interrupt their common trend. Furthermore, a correlation of the Biwabik Iron-formation of the Mesabi Range with the Ironwood Iron-formation of the Gogebic Range has been proposed by many (e.g., Morey 1983) based on close lithologic similarity of the two formations and their enclosing strata. Although some recent interpretations (Morey 1996; Ojakangas et al. 2001) place the Negaunee Iron-formation stratigraphically below the Ironwood–Biwabik–Gunflint units and separated from them by an unconformity, our age

Fig. 4. Concordia diagrams showing SHRIMP U–Pb data for (A) Hemlock volcanic (rhyolite sample MI-99-2) and (B) Tobin Lake quartz monzonite "stitching" pluton (sample MI-99-4). Error ellipses indicate uncertainties at 1σ. Dashed ellipses (showing either excessive discordance or possible inheritance) represent data excluded from weighted mean age calculations. MSWD, mean square of weighted deviates.



determination indicates that the Negaunee Iron-formation is, likewise, correlative with these other three formations. Hence at least four of the major iron formations appear to be correlative.

Tectonic implications

Previous models of diachronous iron-formation deposition inferred the Negaunee Iron-formation and Hemlock Formation were older than the Biwabik–Ironwood units. As discussed earlier in the text, this led to an interpretation that the Negaunee was deposited on a rifted continental margin accompanied by rift volcanism and sea-floor spreading (i.e., rifted Atlantic margin; Ueng et al. 1988; Schulz et al. 1993; Morey 1993; Ojakangas et al. 2001). Rifting and speading in these models occurred after intracratonic sedimentation but before the Ironwood–Biwabik period, which were subsequent products of younger foreland basin development during arc accretion. Based on the regional depositional succession and our new age constraints for the Hemlock Formation (and its correlation with the major iron formations), it is evident that Negaunee and Hemlock deposition was contemporaneous with island-arc volcanism to the south. Therefore, the need for, and validity of, wholesale continental rifting and eventual sea-floor spreading during Menominee Group deposition is called into question. These sequences are now most appropriately interpreted as products of foreland basin sedimentation, which is consistent with that originally outlined by Hoffman (1987), in which iron formations form in foredeeps, particularly along outer ramps, and are approximately synchronous with significant volcanism.

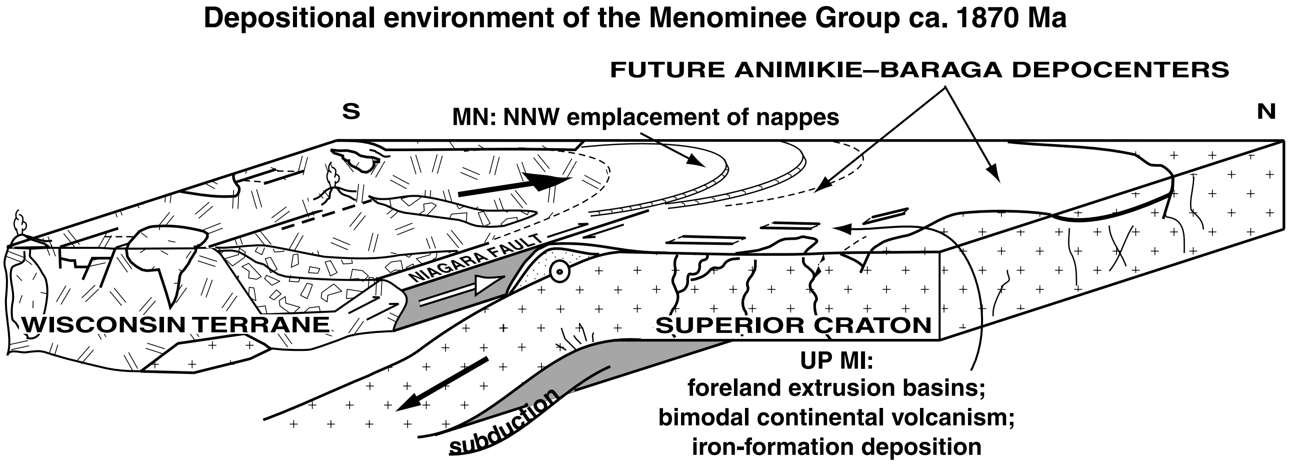
New geochronologic data presented here and in Fralick et al. (1998), albeit limited, combined with reinterpretion of the stratigraphic correlations of the Marquette Range Supergroup, requires a new model for the tectonic setting of MRS deposition. A new model must comply with several geochronologic constraints and fundamental geologic observations:

(1) Menominee volcanism (and iron-formation deposition ca. 1875 Ma) is contemporaneous with volcanism in the Animikie Group north across the entire foreland.
(2) Consistent and correlative lithostratigraphy above and below these dated units, suggests a long period of passive margin development, followed by rapid subsidence due to subduction.
(3) MRS bimodal magmatism is coeval with Wisconsin Magmatic Terrane (island arc) magmatism to the south (ca. 1890–1860 Ma) during subduction.
(4) Minor continental extension, although not rifting, occurred in the subducting plate. We see no evidence for sea-floor spreading during MRS deposition preserved in the rock record of the region, as suggested recently by Ojakangas et al. (2001).

Accommodating these constraints, we outline a revised tectonic framework for Marquette Range Supergroup deposition. We agree with previous interpretations of the lower Chocolay Group, which correlate the glaciogenic deposits with the pre-Nipissing diabase (>2.2 Ga) upper Huronian units (Gair 1981; Young 1983). We then envision a long-lived shallow-marine environment on the margin of the Superior craton, within which were deposited mature sandstones, carbonates, and graywackes (now metamorphosed) of the upper Chocolay Group. This passive-margin suite and the associated overlying unconformity may represent >300 million years of relative tectonic quiescence prior to Penokean subduction and collision. As the northern arc of the Wisconsin Magmatic Terranes formed to the south, beginning about 1890 Ma (Barovich et al. 1989), the depositional setting along the Superior margin became influenced by the initiation of southward directed subduction and arc development.

The previous interpretation that portions, if not all, of the

Fig. 5. Conceptual interpretation of the tectonic environment of deposition for the Menominee Group ca. 1870 Ma. During the early stages of arc convergence, the dominant north-northwest compression in the western orogen creates dextral motion along the Niagara fault zone and minor east–west extension in the eastern orogen.



Menominee Group were deposited in an extensional setting is well supported by discussions presented here. The geochemistry of interbedded rhyolites and the bimodal character of the volcanic units indicate that they were formed in an extending continental environment. Syn-extensional deposition is further supported by the presence of these volcanic suites within minor fault-bounded troughs. These data and observations, combined with the age data presented here and from Fralick et al. (1998), suggest that this extension-related rhyolitic volcanism, occurring north of the Niagara fault zone, was synchronous with compression on or south of the fault zone as the Wisconsin magmatic arcs were developing and being obliquely accreted to the craton margin.

It is important to distinguish this collision-induced extension in the downgoing plate from major extensional rifting and sea-floor spreading during continental breakup, which apparently occurred several hundred million years prior to Menominee Group deposition. The extension we invoke involves minor fault-bounded troughs dissecting only the thin veneer of supracrustal units atop Archean basement. Menominee sedimentation apparently was accompanied by extension generated on the craton margin by otherwise compressive stresses along the subducting margin. The development of continental extension during plate convergence has been documented in several other orogens (Ellis 1986; Ellis and Watkinson 1987; Dostal et al. 1988; Malo et al. 1995). Extension during initial collision is caused by (1) oblique subduction in which deformation in the downgoing footwall produces major extension subparallel to the length of the orogen (Ellis and Watkinson 1987) and (or) (2) collision along an irregular plate margin (Malo et al. 1995). Oblique Penokean subduction, as proposed by Holm et al. (1988) on the basis of structural analysis of Paleoproterozoic supracrustal rocks in Minnesota and aeromagnetic surveys of the Lake Superior Archean margin that strongly suggest a salient–reentrant geometry (Chandler and Southwick, personal communication, 2001), both yield plausible mechanisms for syn-collision extension.

Based on the trend and arcuate shape of thrust structures in Minnesota (cf. Macedo and Marshak 1999), the inherited irregular Archean margin, and the oblique subduction model for the Penokean of Holm et al. (1988), we suggest that during Penokean orogenesis the initial collision-related north-nothwest emplacement of thrust nappes in Minnesota was coeval with early dextral motion along the Niagara fault in Michigan and Wisconsin (Fig. 5). In this transfer zone of oblique motion in Michigan and Wisconsin, a phenomenon related to continental extrusion occurred where crust of irregular shape, thickness, and density collided (i.e., young, warm island-arc rocks converging with rigid Archean basement). Specifically, parts of the southern Superior plate were extruded eastward as the Wisconsin Magmatic Terranes proceeded northwestward. Extrusion produced a complex basin system consisting of several individual sub-basins, orthogonal to collision, and separated by relatively shallow basement blocks (cf. Einsele 2000). These sub-basins exhibit features formed in rift, retroarc, and pull-apart basins: uplifted and downdropped crustal blocks, predominantly clastic sediments derived from nearby highlands, and effects of felsic and alkali basaltic volcanism. The bimodal volcanic sequences of the Hemlock and Emperor formations were probably erupted in these extensional troughs, which also accumulated sediments of the Negaunee and related iron formations and other siliciclastic deposits (now quartzites and slates). Notably, in the eastern Penokean of Michigan and Wisconsin (Fig. 1), extrusion basin occurrence is pronounced due to a >45° trend-change in the Niagara fault.

Synchronous with extrusion basin volcanism ca. 1875 Ma, the Wisconsin Magmatic Terranes maintained growth and convergence, as shown by arc volcanism in the northern arc ca. 1889–1860 Ma. Continental convergence typically leads to eventual deposition of thick marine sediments along the subducting plate margin resulting from rapid subsidence (Einsele 2000) and characterized by deep-water muds and turbidites. Rocks of the Baraga and Animikie Groups (e.g., Michigamme and Virginia formations) are the result of this phase of foredeep sedimentation in the Lake Superior region. These foredeep deposits were derived initially from the northern craton (Archean Superior Province) and later from the approaching highlands of the WMT. The source change is signaled by a Nd signature change seen in the lower to upper Baraga and Animikie Group sediments (Barovich et al. 1989; Hemming et al. 1995).

Deposition of the Marquette Range Supergroup culminated with orogen-normal arc collision in Michigan and Wisconsin at ca. 1875–1860 Ma (Van Schmus 1976, 1980). The crystallization age of the Tobin Lake pluton reported here is similar to other reported post-tectonic stitching plutons in Michigan and northern Wisconsin (Sims et al. 1985). Numerous dates on undeformed granites, some of which pierce the Niagara fault zone, suggest that arc-related collision and accretion with the Superior Province during the Penokean orogeny had ceased by ~1835 Ma. We can now confidently state that convergence during the Penokean orogeny, at least as it is affected by the Wisconsin Magmatic Terrane, had begun by ca. 1875 Ma and was completed by 1835 Ma.

Conclusions

We have attempted to correlate major iron formations of the Marquette Range Supergroup and constrain their age of deposition. Our data support the original interpretations of Hoffman (1987), which suggested contemporaneous deposition of Proterozoic iron formations and indicates that significant amounts of magmatism can be generated within continental crust in compressional foredeeps. Our model invokes oblique Penokean subduction producing fault-bounded sub-basins in the subordinate plate. Further, the model is simpler than those currently advanced, which invoke crustal-scale rifting and sea-floor spreading 10 million years before arc collision (Ueng et al. 1988; Barovich et al. 1989; Schulz et al. 1993; Morey 1993; Ojakangas et al. 2001). The regionally extensive development of rhyolitic sequences and the agreement of our rhyolite age with that in the Gunflint Range of Fralick et al. (1998) in western Ontario demonstrate the considerable extent of extrusion basin growth across the Penokean foreland. We suggest that the ages of the rhyolite units most likely represent the commencement of extrusion basin development, typified by coeval extensional volcanism and sedimentation, and demonstrate that they are comparable with ages of subduction-related arc volcanism to the south.

This model for production of continental extrusion basins is similar to that of the Pannonian Basin of the Carpathians (e.g., Einsele 2000); we refrain from classifying these Penokean features as back-arc basins because of their positioning relative to the then-active arc. We emphasize that in our example, oblique convergence resulted in relatively minor extension of the upper Superior plate and produced only secondary basins with contemporaneous bimodal volcanism. We believe this model best explains the current understanding of geochronologic, structural, and stratigraphic data for the region, and demands further isotopic analyses.

Acknowledgments

This research was supported by funds from the U.S. Geological Survey. G. L. Wortman assisted in sample collection and mineral separation. Discussions with Richard Ojakangas, David Southwick, Glen Morey, Mark Jirsa, Terry Boerboom, and Val Chandler greatly aided in developing the new model proposed here. Constructive comments by Robert Rainbird, Gene LaBerge, and Suzanne Nicholson are appreciated.

References

Barovich, K., Patchett, P.J., Peterman, Z.E., and Sims, P.K. 1989. Nd isotopes and the origin of 1.9–1.7 Ga Penokean continental crust of the Lake Superior region. Geological Society of America Bulletin, **101**: 33–338.

Baedecker, P.A. 1987. Methods for geochemical analysis: U.S.G.S. Bulletin **1770**.

Cambray, F.W., Mancuso, J.J., and Slitor, W. 1991. Detachment faulting and the origin of the asymmetric depositional pattern of the Marquette trough. *In* 37th Annual Institute on Lake Superior Geology, Eau Claire, Wisconsin. pp.17–18.

Cannon, W.F. 1986. Bedrock geologic map of the Iron River 1° × 2° quadrangle, Michigan and Wisconsin: U.S. Geological Survey, Miscellaneous Investigation Series Map I-1360-B, scale 1 : 250 000.

Cannon, W.F., and Gair, J.E. 1970. A revision of stratigraphic nomenclature of middle Precambrian rocks in northern Michigan. Geological Society of America Bulletin, **81**: 2834–2846.

Cannon, W.F., and Klasner, J.S. 1976. Geologic map and geophysical interpretation of the Witch Lake quadrangle, Marquette, Iron, and Baraga Counties, Michigan. U.S. Geological Survey, Miscellaneous Investigations Series, Map I-9897, scale 1 : 62 500.

Clements, J.M., and Smyth, H.L. 1899. The Crystal Falls iron-bearing district of Michigan. U.S. Geological Survey, Monograph 36.

Corfu, F., and Andrews, A. 1986. A U–Pb age for mineralized Nipissing Diabase, Gowganda, Ontario. Canadian Journal of Earth Sciences, **23**: 107–112.

Dostal, J., Wilson, R., and Keppie, J. 1988. Geochemistry of Siluro-Devonian Tobique volcanic belt in northern and central New Brunswick (Canada): tectonic implications. Canadian Journal of Earth Sciences, **26**: 1282–1296.

Eby, G.N. 1992. Chemical subdivision of the A-type granitoids: Petrogenetic and tectonic implications. Geology, **20**: 641–644.

Einsele, G. 2000. Sedimentary basins: evolution, facies, and sedimentary budget. Springer-Verlag, New York.

Ellis, M.A. 1986. Structural morphology and associated strain in the central Cordillera (BC and WA): evidence for oblique tectonics. Geology, **14**: 647–650.

Ellis, M.A., and Watkins, A.J. 1987. Orogen-parallel extension and oblique tectonics: the relation between stretching lineations and relative plate motions. Geology, **15**: 1022–1026.

Fralick, P.W., Kissin, S.A., and Davis, D.W. 1998. The age and provenance of the Gunflint lapilli tuff. *In* 44th Annual Institute on Lake Superior Geology, Minneapolis, Minnesota. pp. 66–67.

Gair, J.E. 1981. Lower Proterozoic glacial deposits of northern Michigan, USA. *In* Earth's pre-Pleistocene glacial record. *Edited by* M.J. Hembrey and W.B. Harland. Cambridge University Press, Cambridge, Chap. F-30, pp. 803–806.

Gair, J.E., and Wier, K.L. 1956. Geology of the Kiernan quad., Iron Co., MI. U.S. Geological Survey, Bulletin 1044.

Goldich, S.S., and Hedge, C.E. 1974. 3.8 Ga granitic gneiss in southwestern Minnesota. Nature, **252**: 467–468.

Hemming, S.R., McLennan, S.M., and Hanson, G.N. 1995. Geochemical and Nd/Pb isotopic evidence for the provenance of the Early Proterozoic Virginia Formation, MN: Implications for the tectonic setting of the Animikie Basin. Journal of Geology, **103**: 147–168.

Hoffman, P. 1987. Early Proterozoic foredeeps, foredeep magmatism and Superior-type iron formations of the Canadian shield. *In* Proterozoic lithospheric evolution. *Edited by* A. Kroner. American Geophysical Union Geodynamics Series, 17, pp. 85–98.

Holm, D.K., and Lux, D. 1996. Core complex model proposed for gneiss dome development during collapse of the Paleoproterozoic Penokean orogen, Minnesota. Geology, **24**: 343–346.

Holm, D.K., Holst, T.B., and Ellis, M.A. 1988. Oblique subduction, footwall deformation, and imbrication: a model for the Penokean orogeny in east-central Minnesota. Geological Society of America Bulletin, **100**: 1811–1818.

James, H.L. 1954. Sedimentary facies of iron formation. Economic Geology, **49**: 235–293.

Karlstrom, K., Ahall, K-I, Harlan, S.S., Williams, M.L., McLelland, J., and Geissman, J.W. 2001. Long-lived (1.8–1.0 Ga) convergent orogen in southern Laurentia, its extensions to Australia and Baltica, and implications for refining Rodinia. Precambrian Research, **111**: 5–30.

Keighin, C.W., Morey, G.B., and Goldich, S.S. 1972. East-central Minnesota. *In* Geology of Minnesota—a centennial volume (MGS). *Edited by* P.K. Sims and G.B. Morey. pp. 240–255.

Kinny, P., Compston, W., and Williams, I. 1991. A reconnaissance ion-probe study of hafnium isotopes in zircons. Geochimica et Cosmochimica Acta, **55**: 849–861.

Klasner, J.S., and Sims, P.K. 1993. Thick-skinned, south-verging backthrusts in the Felch and Calumet troughs area of the Penokean orogen, northern Michigan. U.S. Geological Survey Bulletin 1904-L.

Klasner, J.S., Cannon, W.F., and Ojakangas, R.W 1989. Segment 2—Geology of the Marquette Range Supergroup and the Penokean fold belt in northern Michigan. *In* Early Proterozoic rocks of the Great Lakes region. *Edited by* G.B. Morey. 28th International Geological Congress guidebook T145, American Geophysical Union, pp. 19–31.

Klasner, J.S., LaBerge, G.L., and Cannon, W.F. 1998. Geologic map of the eastern Gogebic iron range, Gogebic County, Michigan: U.S. Geological Survey, Geologic Investigation Series Map I-2606, scale 1 : 24 000.

Klasner, J.S., Ojakangas, R.W., Schulz, K.J., and LaBerge, G.L. 1991. Nature and style of deformation in the foreland of the Early Proterozoic Penokean orogen, northern MI. U.S. Geological Survey Bulletin 1904K.

Leith, C.K., Lund, R.J., and Leith, A. 1935. Precambrian rocks of the Lake Superior region. U.S. Geological Survey, Professional Paper 184.

Macedo, J., and Marshak, S. 1999. Controls on the geometry of fold–thrust belt salients. Geological Society of America Bulletin, **111**: 1808–1822.

Malo, M., Tremblay, A., Kirkwood, D., and Cousineau, P. 1995. Along-strike Acadian structural variations in the Quebec Appalachians: Consequences of a collision along an irregular margin. Tectonics, **14**: 1327–1338.

Morey, G.B. 1983. Lower Proterozoic stratified rocks and the Penokean orogeny in east-central Minnesota. *In* Early Proterozoic geology of the Lake Superior region. *Edited by* L.G. Medaris, Jr. Geological Society of America, Memoir 160, pp. 97–122

Morey, G.B. 1993. Early Proterozoic epicratonic rocks, The Lake Superior region and Trans-Hudson orogen. *In* Precambrian: conterminous United States. *Edited by* J.C. Reed, Jr., M.E. Bickford, R.S. Houston, P.K. Link, D.W. Rankin, P.K. Sims, and W.R. Van Schmus. Geological Society of America, Boulder, Colo., The Geology of North America, Vol. C-2, pp. 47–56

Morey, G.B. 1996. Continental margin assemblage. *In* Archean and Proterozoic geology of the Lake Superior region, U.S.A., 1993. *Edited by* P.K. Sims and L.M.H. Carter. U.S. Geological Survey, Professional Paper, 1556, pp. 30–44.

Morey, G.B., and Southwick, D.L. 1991. Stratigraphic and tectonic framework of the Early Proterozoic Penokean orogen in east-central Minnesota. *In* Precambrian geology of the southern Canadian Shield and the eastern Baltic Shield. *Edited by* R.W. Ojakangas. Minnesota Geological Survey, Circular 34, pp. 42–48.

Morey, G.B., and Van Schmus, W.R. 1988. Correlation chart of Precambrian rocks in the Lake Superior region. U.S. Geological Survey, Professional Paper 1241-F.

Ojakangas, R., Morey, G.B., and Southwick, D.L. 2001. Paleoproterozoic basin development and sedimentation in the Lake Superior region, North America. Sedimentary Geology, **141–142**: 319–341.

Pearce, J.A., Harris, N.B.W., and Twindle, A.G. 1984. Tectonic element discrimination diagrams for the tectonic interpretation of granitic rocks. Journal of Petrology, **25**: 956–983.

Peterman, Z.E., Zartman, R.E., and Sims, P.K. 1986. A protracted Archean history in the Watersmeet gneiss dome. *In* U.S. Geological Survey Bulletin 1655. pp. 51–64.

Schneider, D.A., Holm, D.K., and Lux, D. 1996. On the origin of Early Proterozoic gneiss domes and metamorphic nodes, northern Michigan. Canadian Journal of Earth Sciences, **33**: 1053–1063.

Schulz, K.J., Sims, P.K., and Morey, G.B. 1993. Tectonic synthesis, the Lake Superior region and Trans-Hudson orogen. *In* Precambrian: conterminous United States. *Edited by* J.C. Reed, Jr., M.E. Bickford, R.S. Houston, P.K. Link, D.W. Rankin, P.K. Sims, and W.R. Van Schmus. Geological Society of America, Boulder, Colo., The Geology of North America, Vol. C-2, pp. 60–64.

Sims, P.K 1990. Geologic map of Precambrian rocks of Eau Claire and Green Bay 1° × 2° quadrangle, northern Wisconsin. U.S. Geological Survey, Miscellaneous Investigations Series Map I-1924, scale 1 : 250 000.

Sims, P.K, and Carter, L.M.H. 1993. Archean and Proterozoic geology of the Lake Superior region, USA. U.S. Geological Survey, Profressional Paper 1556.

Sims, P.K., Peterman, Z.E., and Schulz, K.J. 1985. Dunbar gneiss-granitoid dome- Implications for Proterozoic tectonic evolution of northern Wisconsin. Geological Society of America Bulletin, **96**: 1101–1112.

Sims, P.K., Van Schmus, W.R., Schulz, K.J., and Peterman, Z.E. 1989. Tectonostratigraphic evolution of the Early Proterozoic Wisconsin Magmatic Terranes of the Penokean orogen. Canadian Journal of Earth Sciences, **26**: 2145–2158.

Sims, P.K., Schulz, K.J., and Peterman, Z.E. 1992. Geology and geochemistry of Early Proterozoic rocks in the Dunbar area, northeastern Wisconsin. U.S. Geological Survey, Professional Paper 1517.

Sims, P.K., Schulz, K.J., DeWitt, E., and Brasaemle, B. 1993. Petrography and geochemistry of Early Proterozoic granitoid rocks in Wisconsin magmatic terranes of Penokean oeogen, northern Wisconsin — a reconnaissance study. U.S. Geological Survey Bulletin 1904-J.

Stern, R.A. 1997. The GSC Sensitive High Resolution Ion Microprobe (SHRIMP): analytical techniques of zircon U–Th–Pb age determination and performance evaluation. *In* Radiogenic age and isotopic studies, Report 10. Geological Survey of Canada, Current Research 1997-F, pp. 1–31.

Thompson, R.N., Morrison, M.A., Dickin, A.P., and Hendry, G.L. 1983. Continental flood basalt—arachnids rule OK? *In* Continental basalts and mantle xenoliths. *Edited by* C.J. Hawkesworth and M.J. Norry. Shiva Publications, Nantwich, U.K., pp. 158–185.

Trent, V.A. 1973. Geologic map of the Marenisco and Wakefield NE quadrangles, Gogebic County, Michigan: U.S. Geological Survey, Open-file report 73-280, scale 1 : 20 000.

Ueng, W.C., Fox, T.P., Larue, D.K., and Wilband, J.T 1988. Geochemistry and petrogenesis of the Early Proterozoic Hemlock volcanic rocks and the Kiernan sills, southern Lake Superior region. Canadian Journal of Earth Sciences, **25**: 528–546.

Van Hise, C.R., and Leith, C.K. 1911. The geology of the Lake Superior region. U.S. Geological Survey, Monograph 57.

Van Schmus, R. 1976. Early and Middle Proterozoic history of the Great Lakes area, North America. Royal Society of London Philosophical Transactions, **280**: 605–628.

Van Schmus, R. 1980. Chronology of igneous rocks associated with the Penokean orogeny in WI. *In* Selected studies of Archean gneisses and lower Proterozoic rocks, southern Canadian Shield. *Edited by* G. Morey and G. Hanson. Geological Society of America Special Paper 182, pp. 159–168.

Van Schmus, W.R., and Bickford, M.E. 1981. Proterozoic chronology and evolution of the Midcontinent region, North America. *In* Precambrian plate tectonics. *Edited by* A. Kroner. Elsevier, New York, pp. 261–296.

Van Wyck, N., and Johnson, C. 1997. Common Pb, Sm–Nd, and U–Pb constraints on the petrogenesis, crustal architecture, and tectonic setting of the Penokean orogen (Paleoproterozoic) in Wisconsin. Geological Society of America Bulletin, **109**: 799–808.

Young, G.M. 1983. Tectono-sedimentary history of the Early Proterozoic rocks of the northern Great Lakes region. *In* Early Proterozoic geology of the Great lakes region. *Edited by* G. Medaris. Geological Society of America, Memoir 160, pp. 15–32.