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The Penokean orogeny in the Lake Superior region

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Abstract

The Penokean orogeny began at about 1880 Ma when an oceanic arc, now the Pembine–Wausau terrane, collided with the southern margin of the Archean Superior craton marking the end of a period of south-directed subduction. The docking of the buoyant craton to the arc resulted in a subduction jump to the south and development of back-arc extension both in the initial arc and adjacent craton margin to the north. A belt of volcanogenic massive sulfide deposits formed in the extending back-arc rift within the arc. Synchronous extension and subsidence of the Superior craton resulted in a broad shallow sea characterized by volcanic grabens (Menominee Group in northern Michigan). The classic Lake Superior banded iron-formations, including those in the Marquette, Gogebic, Mesabi and Gunflint Iron Ranges, formed in that sea. The newly established subduction zone caused continued arc volcanism until about 1850 Ma when a fragment of Archean crust, now the basement of the Marshfield terrane, arrived at the subduction zone. The convergence of Archean blocks of the Superior and Marshfield cratons resulted in the major contractional phase of the Penokean orogeny. Rocks of the Pembine–Wausau arc were thrust northward onto the Superior craton causing subsidence of a foreland basin in which sedimentation began at about 1850 Ma in the south (Baraga Group rocks) and 1835 Ma in the north (Rove and Virginia Formations). A thick succession of arc-derived turbidites constitutes most of the foreland basin-fill along with lesser volcanic rocks. In the southern fold and thrust belt tectonic thickening resulted in high-grade metamorphism of the sediments by 1830 Ma. At this same time, a suite of post-tectonic plutons intruded the deformed sedimentary sequence and accreted arc terranes marking the end of the Penokean orogeny. The Penokean orogen was strongly overprinted by younger tectonic and thermal events, some of which were previously ascribed to the Penokean. Principal among these was a period of vertical faulting in the Archean basement and overlying Paleoproterozoic strata. This deformation is now known to have post-dated the terminal Penokean plutons by at least several tens of millions of years. Evidence of the Penokean orogen is now largely confined to the Lake Superior region. Comparisons with more recent orogens formed by similar plate tectonic processes implies that significant parts of a once more extensive Penokean orogen have been removed or overprinted by younger tectonic events.

Keywords: Paleoproterozoic tectonics; Penokean orogeny; Stratigraphy; Structure; Magmatism; Lake Superior region

1. Introduction

Laurentia, the North American craton, consists of a network of Paleoproterozoic orogenic belts that surround Archean crustal provinces. Many of the Paleoproterozoic orogenic belts appear to be collisional zones between formerly independent Archean microcontinents, whereas others are zones of lateral accretion of mainly juvenile island arcs and associated intraoceanic deposits (Hoffman, 1988). The southern margin of Laurentia, in particular, appears to have been a long-lived (1.9–1.0 Ga) convergent orogen that records a complex system of plate margin and island arc activity that resulted in episodic southward continental growth (Windley, 1992; Rivers and Corrigan, 2000; Karlstrom et al., 2001) in a manner analogous to Cordilleran-type plate margins.

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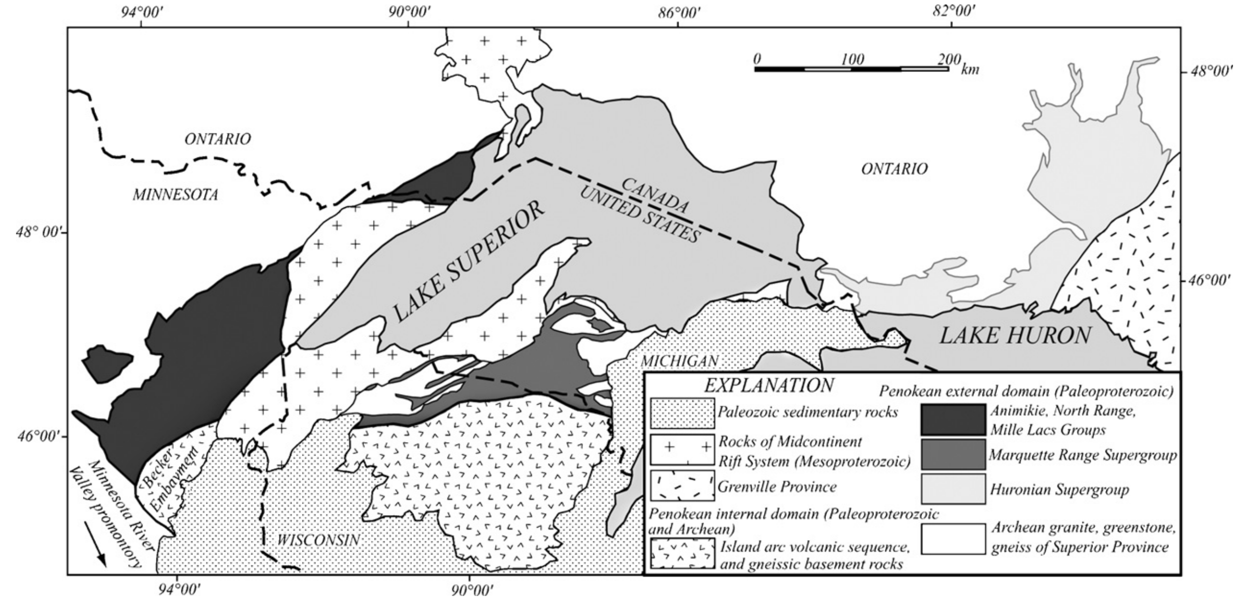

Fig. 1. Generalized geologic map of the Lake Superior–Lake Huron region showing the distribution of Paleoproterozoic stratigraphic sequences of the external continental margin domain and the magmatic rocks of the internal domain of the Penokean orogen.

The Penokean orogen is the oldest Paleoproterozoic accretionary orogen along the southern margin of Laurentia and is interpreted to have developed in an embayment along the southern margin of the Archean Superior craton (Hoffman, 1988; Sims, 1996). It has generally been extended from Minnesota eastward to the Grenville orogen in the Lake Huron region (Sims, 1996) and southwestward to the Central Plains orogen (Sims and Peterman, 1986; Hoffman, 1988). Where exposed in the Lake Superior region, the Penokean orogen consists of a southern internal and a northern external domain (Fig. 1). The external domain consists of a deformed continental margin-foreland basin sequence overlying an Archean basement. The supracrustal rocks include the Paleoproterozoic Marquette Range Supergroup in Michigan and Wisconsin (Cannon and Gair, 1970) and the Animikie, North Range, and Mille Lacs Groups in Minnesota (Morey, 1973). The internal domain consists of an assemblage of island arc terranes, generally referred to as the Wisconsin magmatic terranes (Sims et al., 1989; Sims and Schulz, 1996). These arc terranes mostly consist of Paleoproterozoic tholeiitic and calc-alkaline volcanic rocks and coeval calc-alkaline plutonic rocks, but include Archean basement in the southern (Marshfield) terrane (Sims et al., 1989). The two domains are separated by the broadly accurate, convex-northward Niagara fault (suture) zone in northern Michigan and Wisconsin (Fig. 2). In Minnesota, the suture appears to have been destroyed by geon 17 granites. The Penokean orogeny involved northward-directed thrusting and related folding of the supracrustal rocks of the northern domain resulting from the collision of the southern Wisconsin magmatic terranes at about 1.88–1.85 Ga (Sims et al., 1985, 1989; Sims, 1996).

1.1. Historical perspective on the Penokean orogeny

The term "Penokean orogeny" was introduced by Blackwelder (1914) for a period of mountain building that he considered as post-Keweenawan (Mesoproterozoic) based on inferred relations in the Penokee Range (now commonly called the western Gogebic Range) in northern Wisconsin. This interpretation was based on the long-held but erroneous concept that there had been no major orogeny during the interval from the end of the Archean through to the Mesoproterozoic (Keweenawan). It was not until the mid-1950s that it was recognized that an angular unconformity and discontinuities in metamorphic grade separated the Paleoproterozoic supracrustal rocks from the Mesoproterozoic Keweenawan rocks in the Lake Superior region (Marsden, 1955; James, 1958). Goldich et al. (1961) recognized the wide-spread occurrence of post-Animikie–pre-Keweenawan deformation in the Lake Superior region and adopted the term "Penokean" for this orogenic event.

Cannon (1973) synthesized information on the Penokean orogeny based on the extensive mapping by the U.S. Geological Survey of Paleoproterozoic rocks in the iron ranges in northern Michigan. He recog-

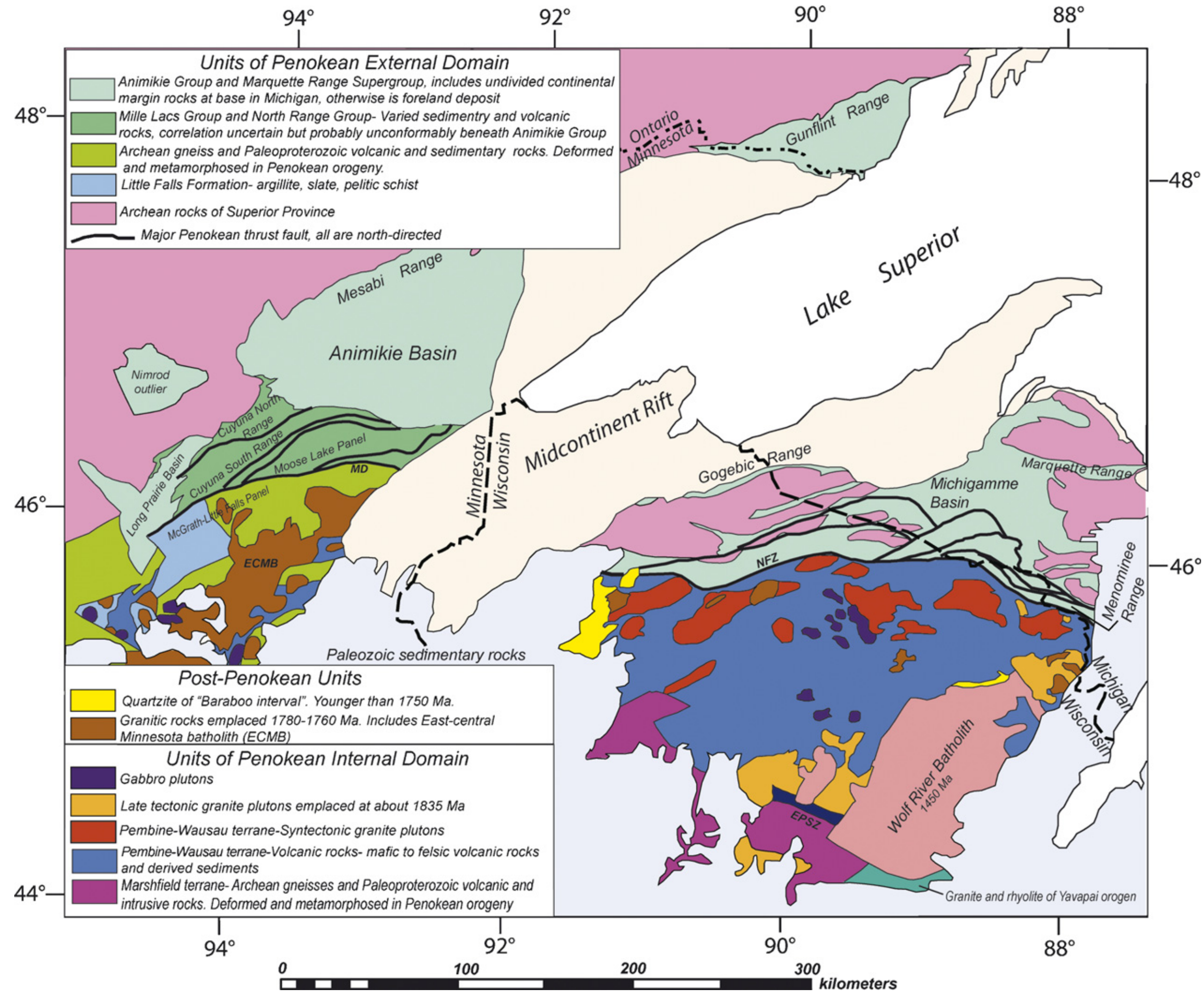

Fig. 2. Generalized geologic map of the Penokean orogen showing features referred to in the text. *Abbreviations*: ECMB, East-central Minnesota batholith; EPSZ, Eau Pleine shear zone; MD, Malmo discontinuity; NFZ, Niagara fault zone.

nized a wide divergence of trends of major (first-order) folds in the rocks of the Marquette Range Supergroup and a lack of Paleoproterozoic penetrative deformation in Archean basement rocks. He proposed a tectonic model involving an early phase of regional gravity sliding of Paleoproterozoic strata that were detached from Archean basement followed by vertical uplift of Archean basement blocks. The latter vertical uplift phase was responsible for the majority of the current map pattern of domes of Archean basement and keels of Paleoproterozoic strata (Cannon, 1973).

Most workers through about the mid-1970s interpreted the Paleoproterozoic history in the Lake Superior region in terms of intracratonic basin development followed by intracratonic deformation related to vertical crustal movements (Cannon, 1973; Morey, 1973; Sims, 1976). Van Schmus (1976) was the first to suggest that the rocks and structures formed during the Penokean orogeny were similar to those found in more recent orogenic belts formed by plate tectonic processes. This concept has since emerged as the general consensus view (Cambray, 1978, 1984; Larue, 1983; LaBerge et al., 1984; Schulz et al., 1993; Sims, 1987; Hoffman, 1987, 1988; Southwick et al., 1988) from: (1) further documentation of the similarity between the Paleoproterozoic supracrustal rocks of the external domain and those of recent rifted, passive margins and foreland basins (Larue, 1981a,b; Larue and Sloss, 1980; Hoffman, 1987; Schulz et al., 1993); (2) documentation of structures within the Paleoproterozoic supracrustal rocks of the external domain including folds and thrust faults similar to those in recent collisional fold and thrust belts (Holst, 1984, 1991; Holm et al., 1988; Klasner et al., 1991; Southwick and Morey, 1991; Gregg, 1993); and (3) documentation of dominantly calc-alkaline volcanic and plutonic rocks of island-arc affinity in the juxtaposed Paleoproterozoic Wisconsin magmatic terranes (Van Schmus, 1976; Greenberg and Brown, 1983; Sims et al., 1985, 1989, 1993). The results of studies of the Penokean orogen in the Lake Superior region through the early 1990s were summarized by Sims and Carter (1996).

Van Schmus (1976) found that the Penokean events ranged in age from about 1890–1830 Ma; the end of the Penokean being marked by a suite of post-tectonic granite plutons. Recent geochronologic studies have documented a rich history of metamorphic, tectonic, and igneous events during the succeeding geon. In this paper, however, we retain Van Schmus' temporal definition for the Penokean and consider the slightly younger events as post-Penokean. We consider the post-tectonic 1830 plutonism to constitute an unequivocal upper bound on the Penokean orogeny and to record at least a brief cessation in tectonism prior to renewed activity in early geon 17.

1.2. The Penokean orogeny: a new synthesis

In this paper we summarize a wealth of previously published information on the lithologic and tectonic evolution of the Penokean orogen and also discuss recently published geochronologic data for rock-forming, metamorphic, and tectonic events, both in the external domain (foreland basin) and in the internal arc terranes. The increasingly well dated history of these two terranes, taken together, allows us to propose a model that is more detailed and comprehensive and links foreland events to those in the marginal accretion zones. We also discuss the implications of age determinations derived from within the orogen that indicate that there was a previously under-appreciated tectonic overprint on the Penokean orogen from both the Yavapai and Mazatzal orogenies. Many features previously ascribed to Penokean tectonism are now reasonably interpreted to be a result of these younger orogenies. Although our purpose is not to discuss these younger events in detail (see other papers in this volume for more detailed discussions) it is important to properly distinguish those events from those of truly Penokean-age in order to correctly define the Penokean history of the region.

2. The Penokean orogen

2.1. External domain—the northern continental margin sedimentary basin

The external domain consists of a suite of sedimentary and lesser volcanic rocks that were deposited in both the rifting phase of pre-Penokean ocean opening and the foreland basin phase of Penokean arc accretion (see Fig. 3 for stratigraphic units and their inferred correlations within the external domain). The older part of the sequence, the Chocolay Group, consisting of mature quartzites, carbonates and sporadically preserved glaciogenic deposits, is now known to have been deposited between 2.3 and 2.2 Ga and thus to correlate with lithologically similar units in the upper part of the Huronian Supergroup in Ontario (Vallini et al., 2006). In Ontario, these upper Huronian rocks have been interpreted as a continental margin sequence (Young and Nesbitt, 1985) that was preceded by the deposition of rift-related volcanic and sedimentary rocks. In Michigan, the Chocolay Group was interpreted to have been deposited in rift or sag basins and intervening platforms on an evolving continental margin (Larue and Sloss, 1980). The bulk of the continental margin sequence in the Lake Superior region was deposited in a foreland basin north of the northward advancing accreting arcs. The sedimentary and interbedded volcanic rocks are the products of the dynamic interaction of subsidence driven by the tectonic loading of the continental margin and rapid uplift and erosion as the foreland sediments were incorporated into the advancing fold and thrust belt. The foreland basin assemblage includes rocks of the classic iron ranges for which the Lake Superior region is well known. All the rocks of the continental margin assemblage were deposited on Archean rocks on the southernmost extent of the Superior craton. The Archean rocks range in age from about 3.5–2.6 Ga and range from granites and gneisses to volcanic and related volcanogenic sedimentary rocks (Sims and Carter, 1996).

In the following paragraphs we present a general outline of the lithologic and stratigraphic features of the continental margin. A general stratigraphic framework has been long established (Fig. 3), and readers are referred to recent summaries for details (Morey, 1996; Ojakangas et al., 2001). Recent geochronologic advances have clarified some inter-district correlations and provide an improved absolute time frame in which various sequences can be assigned tectonic position. This tectono-stratigraphic sequence is outlined in Fig. 4 and the significance of recent geochronologic constraints is discussed in the following paragraphs.

2.1.1. Michigan and Wisconsin

Deposition and deformation of the continental margin assemblage is best known in the various iron-ranges where outcrops are generally best and where a century and a half of mining and geologic studies provide an extensive base of geologic data. Rocks of the Choco-

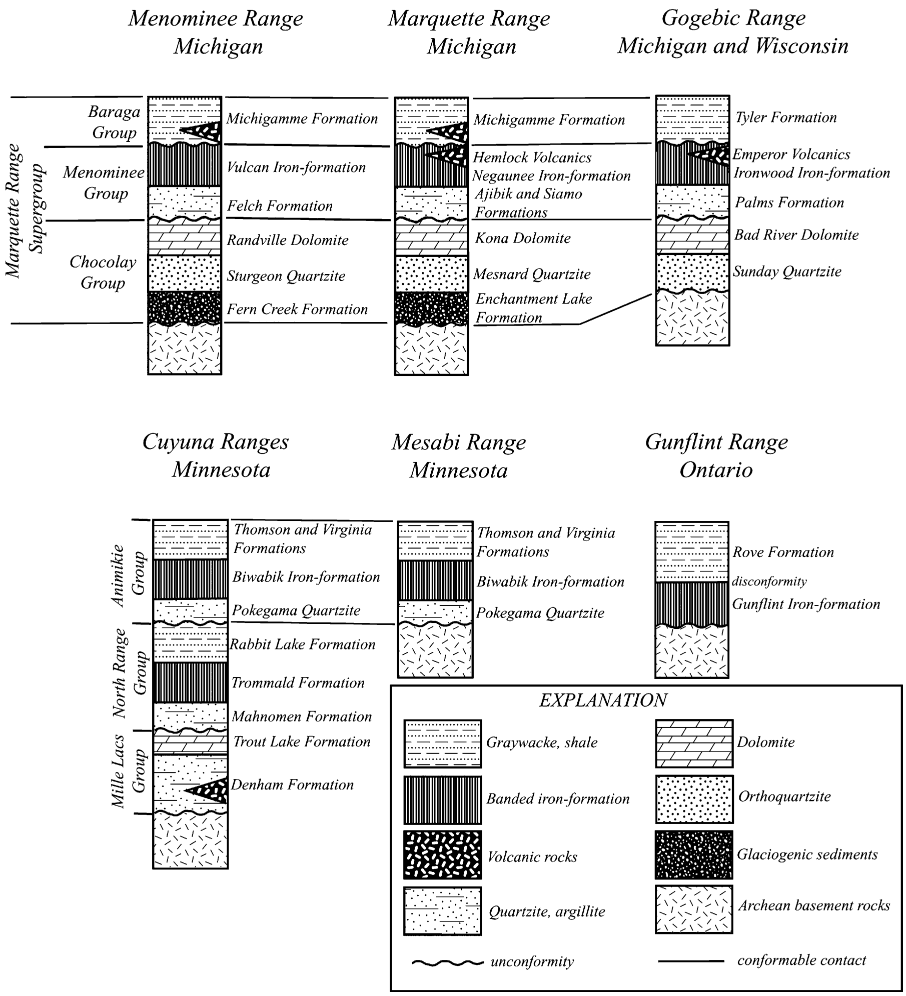

Fig. 3. Stratigraphy of the major iron ranges of the Lake Superior region generalized and modified from Ojakangas et al. (2001). See Fig. 2 for locations of the iron ranges.

lay Group are preserved only sporadically in this region, probably as a result of as much as 300 million years of erosion between their deposition and deposition of the overlying Menominee Group (Fig. 4). Chocolay deposition has not been dated directly, but is bracketed between about 2.3 Ga, the age of the youngest detrital zircons known in clastic rocks, and 2.2 Ga, the age of hydrothermal xenotime cement in these same rocks (Vallini et al., 2006). The most recent thorough studies of the Chocolay Group (Larue and Sloss, 1980; Larue, 1981a) suggested a depositional setting in broad rift basins and intervening platforms on an evolving continental margin. These rocks, thus, may record the earliest subsidence related to continental breakup. If so, there is a remarkably long hiatus (as much as 300 m.y.) between breakup and subsequent arc accretion, which implies the development of a very extensive ocean to the south of the region.

The earliest imprint of Penokean events was regional subsidence marked by onset of the widespread deposition of the shallow marine clastic rocks of the basal Menominee Group followed by the deposition of iron-formation, mostly in shallow water, in part above wave-base, environments (Larue, 1981b) (Figs. 3 and 4). Iron-formation deposition appears to have occurred in somewhat contrasting settings varying from broad stable platforms such as in most of the Gogebic Range, to subsiding grabens, such as in the Marquette Range and eastern part of the Gogebic Range. The iron-formation passes laterally into volcanic rocks, which have yielded a U/Pb zircon age of 1874 ± 9 Ma (Schneider et al., 2002). The similarity of this age to ages for volcanic and related intrusive rocks as well as massive sulfide deposits in the accreted Pembine–Wausau magmatic terrane clearly establishes the contemporaneousness of

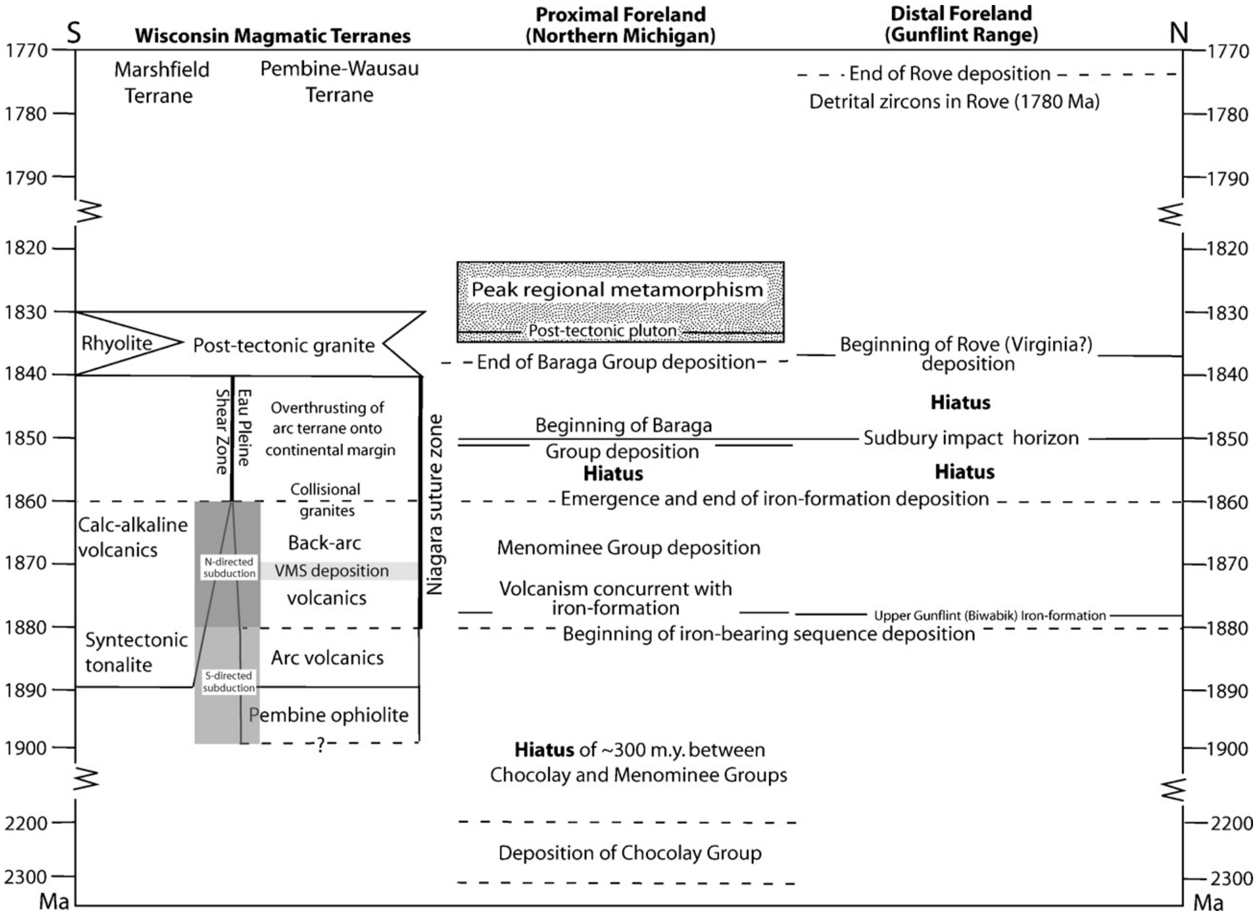

Fig. 4. Comparison of the sequence of Penokean events in the magmatic arc terranes, the proximal foreland, and distal foreland. Solid lines are based on radiometric ages. Dashed lines are estimated times between radiometrically bracketed events.

magmatic activity in the external and internal domains of the Penokean orogen. The grabens into which some of the iron-formation was deposited have been interpreted to be second-order extensional basins resulting from oblique collision and northward accretion of the Wisconsin magmatic terranes (Schneider et al., 2002), but we propose an alternative explanation below involving back-arc spreading. The volcanic rocks, mostly tholeiitic basalt and lesser rhyolite, interfinger with the iron-formation and have a within-plate chemical signature (Schneider et al., 2002) indicating that these volcanic rocks are also deposited in extensional basins within the cratonic margin.

Iron-formation deposition was terminated by uplift, followed by a period of subaerial weathering and erosion. In places, the Menominee Group is now absent, partly a result of that erosion, although the original extent of that group is not known. Its absence may also result, in part, from non-deposition. Neither the duration nor the tectonic significance of this uplift is entirely clear. A hiatus that lasted on the order of 10 m.y. is suggested on Fig. 4. We suggest that the uplift might record the passage of a flexural bulge formed at the onset of major overthrusting of arc rocks and the loading of the southern edge of the craton.

Beginning at about 1850 Ma the area was resubmerged and rapidly evolved into a deep-water turbidite basin into which the Baraga Group was deposited. That group consists of thin basal clastic rocks deposited in shallow water environments, local iron-formations, and black pyritic shale followed by a very thick succession of turbidites and lesser volumes of mafic volcanic rocks (Fig. 3) (Ojakangas, 1994). In general, volcanic rocks are more abundant in the southern, more proximal, parts of the basin, but a within-plate chemical signature (Schulz, unpublished data) indicates that they were deposited in the foreland basin and are not overthrust arc rocks. Both paleocurrent and isotopic data indicate that most of the sediments were derived from the synchronous arc terranes to the south (Barovich et al., 1989; Ojakangas, 1994). The beginning of the Baraga Group deposition has not been well constrained. Recently, a layer of ejecta formed during the 1850 Ma Sudbury meteor impact was documented in the Gunflint and Mesabi Ranges in Ontario and northern Minnesota. The layer lies between the top of the iron-formation and the base of the Rove Formation (Addison et al., 2005). We have also tentatively identified this layer in Michigan, where it occurs near the base of the Baraga Group indicating that deposition began at about 1850 Ma (Cannon et al., 2006).

The thickness of the Baraga Group turbidite sequence cannot be constrained because of poor outcrops, complex folding, and the lack of marker horizons. A minimum thickness of several kilometers is inferred from areas of uniform facing directions and similar bedding orientations on the limbs of large folds. The onset of turbidite deposition appears to mark the arrival of this part of the continental margin at a very proximal position relative to the accreting arc terranes. Deposition was rapid and was soon followed by northward-directed compressional deformation and high-grade metamorphism. Most foreland deformation and metamorphism of the Penokean orogeny must have occurred in the 15 m.y. time span from the start of the deposition of the Baraga Group until the emplacement of post-tectonic granite plutons at about 1833 Ma (Schneider et al., 2002). Peak metamorphic temperatures were reached at about 1830 Ma (Schneider et al., 2004) and at a depth of about 15 km (Attoh and Klasner, 1989). We interpret these data to indicate that the foreland, which began as a broad, shallow basin having local extensional basins that host the major iron-formations, evolved into a rapidly subsiding turbidite basin that was incorporated into a northward-advancing fold and thrust belt soon after deposition. At least 15 km of foreland rocks were tectonically stacked by 1830 Ma.

2.1.2. *Mesabi and Gunflint ranges*

The most distally preserved rocks of the Penokean foreland basin are the rocks of the Animikie Group that are preserved in the Mesabi and Gunflint Ranges (Fig. 2). Although these two ranges are not physically contiguous at the present surface, they are separated by a gap of only 100 km underlain by younger gabbro intrusions. The equivalence of the two ranges and their major stratigraphic units has never been seriously questioned, although different stratigraphic names were applied to equivalent units in each range (Fig. 3). There are some stratigraphic similarities to the proximal foreland of Michigan; however, several significant differences are discussed below. Rocks equivalent to the Huronian Supergroup or Chocolay Group are not known. The oldest Paleoproterozoic rocks are part of the iron-bearing sequence (lower Animikie Group) and lie unconformably on Archean rocks (Figs. 3 and 4). Sedimentation began with a thin unit of quartz sand in a tidal environment (Ojakangas et al., 2001) and quickly evolved into a widespread and laterally uniform iron-formation. The internal stratigraphy of the iron-formation, which reflects fluctuations in water depth, can be traced for the entire strike length of the Mesabi Range. The iron-formation is overlain by clastic rocks of the Rove Formation in the Gunflint Range and by the equivalent Virginia Formation in the Mesabi Range. Both units vary from fine- to coarse-grained shales and lithic arenites.

Until recently, the iron-formations and overlying clastic rocks were thought to be a continuous depositional sequence that recorded incursion of clastic-dominated sediments into the previous sediment-starved iron-formation basin. However, recent geochronologic results suggest that in the Gunflint Range a previously unrecognized disconformity lies beneath the Rove Formation and that a significant hiatus separates the Rove and Gunflint Formations. A volcanic ash layer near the top of the Gunflint Iron-formation has yielded a U/Pb zircon age of 1878 ± 1.3 Ma (Fralick et al., 2002), and an ash layer near the base of the Rove Formation yielded a U/Pb zircon age of 1836 ± 5 Ma (Addison et al., 2005). These two layers are separated stratigraphically by about 110 m of rocks, mostly of the upper part of the Gunflint Iron-formation, whereas the temporal separation is about 40 m.y. Deposition of the 1850 Ma Sudbury ejecta, which separates the Gunflint and Rove, was interpreted to have been subaerial (Addison et al., 2005). Thus, there is growing evidence for a previously unrecognized period of emergence between deposition of iron-formation and the overlying clastic rocks. Whether this emergence extended southwest into the Mesabi Range is unresolved at this time. This hiatus is correlative with that in Michigan between the iron-bearing Menominee Group and the overlying Baraga Group but appears to have outlasted that hiatus by as much as 15 m.y. (Fig. 4). This illustrates a significant difference in depositional history between the proximal and distal parts of the foreland basin. Although the deposition of the iron-formations in both settings was at least partly correlative, indicating that initial foreland subsidence was widespread and uniform, subsequent events were diachronous. Rove deposition may have been much longer than previously thought. Detrital zircons with U/Pb ages of about 1780 Ma were found in a sandstone bed about 400 m above the base of the Rove (Heaman and Easton, 2005) indicating that Rove deposition outlasted not only sedimentation of

Baraga Group rocks in Michigan, but also outlasted Penokean deformation and metamorphism in the more proximal parts of the orogen (Fig. 4).

Poorly known extensions of the rocks of the Animikie Group are known in the Nimrod outlier and Long Prairie basin to the south and southwest of the main Animikie basin (Fig. 2). The iron-bearing part of the Animikie Group is not known in these basins, which may contain rocks entirely equivalent to the Virginia and Rove Formations (Southwick et al., 1988). The apparent angular unconformity between these rocks and the underlying deformed rocks of the foreland is discussed below.

2.1.3. Proximal fold and thrust belt of Minnesota

Rocks of the proximal fold and thrust belt include those of the Cuyuna North Range, Cuyuna South Range, and Moose Lake and McGrath-Little Falls panels (Fig. 2) as described by Southwick et al. (1988). These rocks constitute the Mille Lacs and North Range Groups (Fig. 3), a diverse suite of rocks whose age and correlation is still uncertain. The older of these, the Mille Lacs Group, contains local units of both orthoquartzite and carbonate, rock types generally found within the Chocolay Group in Michigan and in the Huronian Supergroup in Ontario. As a result, it has commonly been considered to be roughly correlative with them and have an age of approximately 2.2–2.3 Ga. If this correlation is correct, then the Mille Lacs Group was deposited during the early continental rifting and breakup of the Superior craton. The major part of the Mille Lacs Group consists of pelitic rocks at variable metamorphic grades, mafic volcanic rocks and related mafic intrusions, and minor iron-formation, all of which might have been deposited in a rift. Alternatively, the overall character of these rocks, particularly the southward increase in volcanic components, is quite similar to that of the Menominee Group in the tectonically comparable proximal foreland belt of Michigan and Wisconsin where foreland basin deposits have been thrust northward and intensely folded. Intense deformation, as well as sparse outcrops, has hindered a more detailed definition of the age and tectonic affinities of the rocks in Minnesota. The North Range Group lies unconformably on the Mille Lacs Group and is restricted to the Cuyuna North Range. It consists of a sequence of shale and graywacke that has a medial iron-formation, the Trommald Formation (Fig. 3), which was the iron-bearing unit from which the Cuyuna Range iron ores were mined. The Trommald was long considered to be the equivalent of the Biwabik Iron-formation of the nearby Mesabi Range, but later interpretations suggest that the Animikie Group unconformably overlies the North Range Group (Chandler, 1993). The current stratigraphic interpretation was summarized by Ojakangas et al. (2001). Within the McGrath-Little Falls panel, the Little Falls Formation consists mostly of amphibolite facies metapelites the correlation and tectonic affinities of which are poorly known (Southwick et al., 1988).

2.1.4. Synopsis of foreland sedimentation

The sedimentary sequence preserved in the Penokean foreland records a cycle of continental rifting and ocean opening, followed by arc collision, back-arc and foredeep sedimentation and the nearly coeval incorporation into a foreland fold and thrust belt. The oldest sedimentary rocks are the Chocolay Group in Michigan–Wisconsin and probably at least part of the Mille Lacs Group in Minnesota, which consist of widely scattered erosional remnants of an originally widespread terrestrial to shallow marine sequence containing orthoquartzite and carbonate rocks and, locally, glaciogenic deposits. These are correlative with the upper part (Cobalt Group) of the much better preserved Huronian Supergroup north of Lake Huron. The Huronian contains basal units as old as 2450 Ma that were deposited in rift basins during early stages of continental extension (Heaman, 1997). These old rift rocks have not been identified in the Lake Superior region, where the oldest rocks are glaciogenic units whose maximum age is about 2.3 Ga based on the age of contained detrital zircons (Vallini et al., 2006). The depositional setting of the Chocolay Group, however, was interpreted to be broad rift basins on an evolving continental margin (Larue and Sloss, 1980). Thus, early phases of breakup may have been as old 2.3 Ga. The Huronian and Chocolay sedimentation was complete by 2.2 Ga, the age of widespread mafic intrusions in the Huronian and the age of hydrothermal xenotime cements in the Chocolay Group (Vallini et al., 2006). The Fort Frances (Kenora-Kabetogama) dike swarm in Minnesota radiates from the general area of the embayment in the Superior margin (Becker embayment; Fig. 5). These roughly 2.1 Ga dikes have trace element and Nd isotopic compositions that are characteristic of asthenosphere-derived melts (Buchan et al., 1996; Schmitz et al., 1995; Wirth and Vervoort, 1995). Their emplacement may mark the final breakup of the Superior craton (Kenorland continent) and the onset of a passive margin.

An enigma of the Lake Superior region is the absence of any sedimentary rocks that can be ascribed to deposition on that long-lived continental margin; the next youngest sedimentary units being the iron-bearing sequences that marked the onset of arc accretion at about 1880 Ma (see discussion below). More than 300 m.y. separate the Huronian and equivalent Lake Superior rocks

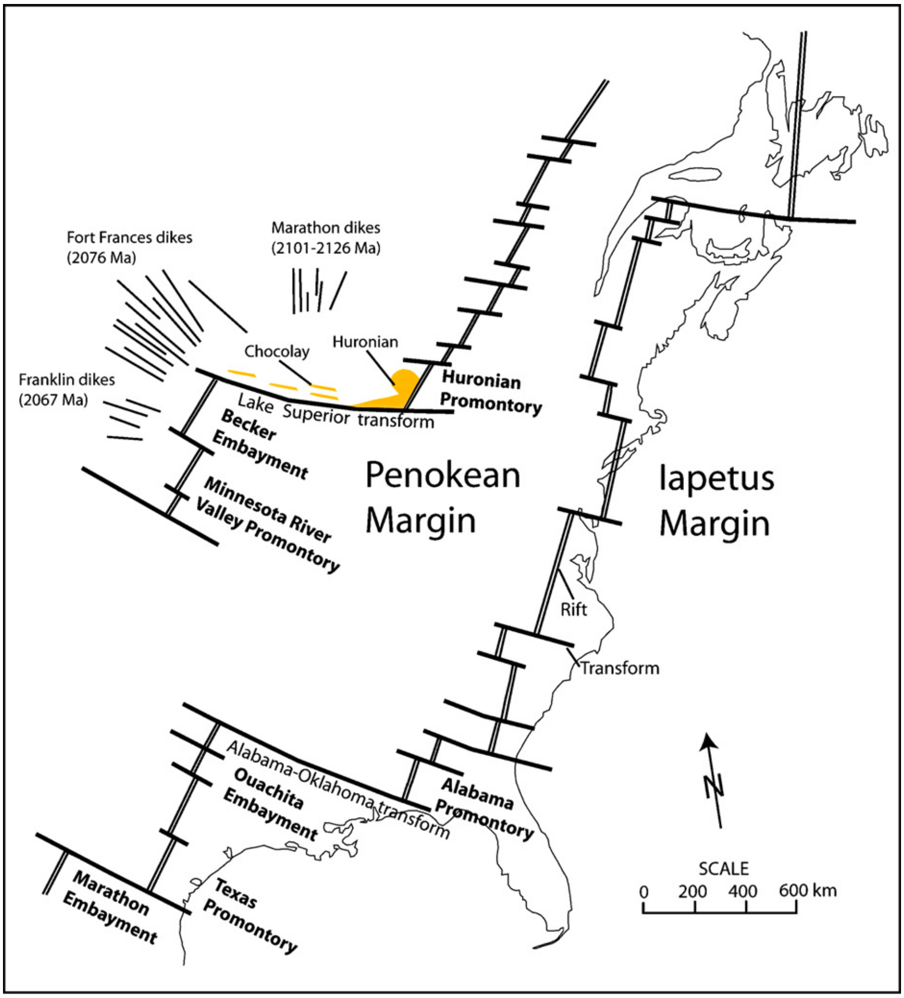

Fig. 5. Schematic map showing the interpreted Paleoproterozoic Penokean continental margin bounded by rift segments, transform faults, and radiating dike swarms (present day coordinates) in comparison with the similarly interpreted late Precambrian–early Paleozoic (Iapetus) eastern North American continental margin (adopted after Thomas, 2006).

from younger deposits of the Penokean external domain (foreland basin). This period, in which no record of sedimentation is known, must have been an extended period of ocean opening and closing prior to the first arc accretion and the onset of Penokean contractional deformation. However, the lack of any significant structural discordance between the Chocolay and Menominee Groups in Michigan indicates that the Superior margin during this period likewise was devoid of significant tectonic deformation. This suggests that the southern Superior craton margin was maintained as an elevated passive margin until the onset of the Penokean orogeny.

Detailed analysis of the Paleoproterozoic rifting history of the southern Superior craton margin is beyond the scope of this paper; however, we suggest that the lack of deposition on this long-lived passive continental margin may relate to the nature of rifting and formation of the embayment in which the Penokean orogen evolved. Although many continental margins experience major subsidence following continental break-up and the start of sea floor spreading, usually with formation of thick sedimentary piles (Bott, 1992), others are characterized by little or no subsidence and elevated plateaus, often with pronounced escarpments (Japsen et al., 2006). Elevated passive margins also are generally characterized by relatively narrow transitions from full-thickness continental crust to oceanic crust and have no or limited postrift (passive margin) sedimentary sequences (Lister et al., 1986; Thomas, 1993; Lorenzo, 1997). Elevated passive continental margins have been attributed to: (1) formation as the upper plate of a low-angle detachment extensional system (Lister et al., 1986, 1991); (2) magmatic underplating at volcanic passive margins (Lister et al., 1991); and (3) transform faulting (Thomas, 1991; Lorenzo, 1997).

Upper plate passive margins, characterized by thick continental crust, narrow continental shelves and thin sedimentary cover, generally form promontories along continental margins that are complementary to opposing lower plate passive margin embayments (Hansen et al., 1993; Thomas, 1993). An upper plate model would, therefore, not appear to be applicable to the formation of the embayment in which the Penokean orogen evolved. An upper plate margin model may apply, however, to the formation of the promontory occupied by the Minnesota River Valley terrane and may account for the lack of Paleoproterozoic sediments along its southern extent in northern Iowa and southern Minnesota (Chandler et al., 2007).

There are no thick sequences of volcanic rocks of appropriate age along the Superior craton margin in the Lake Superior region to indicate that it evolved as a volcanic passive margin. There are, however, basaltic dike swarms, including the Marathon dikes north of Lake Superior (∼2126 Ma, Halls et al., 2006), Fort Frances (Kenora-Kabetogama) dikes in northwest Minnesota-southwest Ontario (∼2076 Ma, Buchan et al., 1996), and Franklin dikes in the Minnesota River valley (∼2067 Ma, Schmitz et al., 2006), which suggest that magmatic underplating may have occurred (Fig. 5). The trends of the Marathon and Fort Frances dikes, taken as a whole, define a radiating swarm with a fan angle of 140° that converges to a focal region in central-southern Wisconsin (Halls et al., 2006). Halls et al. (2006) have suggested that the dike swarms relate to a Paleoproterozoic mantle plume in the Lake Superior region centered approximately beneath the area of the marginal embayment. In this model, the embayment would have formed at the intersection of two successful arms of a three-armed radial, plume-related rift (rift-rift-rift triple junction) (Burke and Dewey, 1973).

Continental-ocean fracture zones are the fossil transform offsets located along passive continental margins and have been proposed to be responsible for the kinked, zig-zag appearance of some present-day continental margins (Thomas, 1977; Lorenzo, 1997). Continental margin segments that form as a result of transform faulting are characterized by a relatively abrupt transition from full-thickness continental crust to oceanic crust, local fault-bounded basins parallel to the transform fault, and thin or absent post-rift deposits (Thomas, 1991, 1993). In the initiation of a transform margin, rifting at a point on the continental crust experiences shearing and the formation of a narrow graben along which the transform fault will eventually rupture (Lorenzo, 1997). As the end of the spreading center passes along the margin, it causes heating and thermal uplift of the adjacent continent. This heating can result in related magmatism while the thermal uplift results in denudation. Once the spreading center has passed, shearing ceases and the margin is subjected only to thermal subsidence (Lorenzo, 1997). Because transform margins have steep continental slopes, shelf sediments tend to pass through the continental slope directly to the abyssal plain (Basile et al., 2005). We suggest that the geometry of the Paleoproterozoic Superior craton margin was generally similar to that presented for the Appalachian–Ouachita (Iapetus) rifted margin of the southern United States (Thomas, 1991, 1993, 2006). The embayment in the Superior margin (Becker embayment, Chandler et al., 2007) has a configuration similar to that of the Ouachita embayment (Fig. 5), which has been attributed to formation by transform rifting along the Alabama–Oklahoma transform (Thomas, 1991).

We conclude that the embayment in which the Penokean orogen evolved may have formed as the result of rifting at a plume-generated triple junction and/or transform faulting. Either processes or their combination would have resulted in an elevated passive margin characterized by thick continental crust with a potentially narrow continental shelf and little or no passive margin sediments. Further work, however, is required to detail the Paleoproterozoic rifting history of the Superior craton margin.

Deposition on the Superior margin began with a transgressive sequence of basal siliciclastic rocks and pelites followed by major iron-formations in an evolving back-arc basin. The Gunflint Iron-formation in Ontario, the most distal preserved part of the basin, and the iron-formations in Michigan, in the proximal part of the basin, have no distinguishable difference in age based on recent precise zircon geochronology. Thus, the onset of sedimentation appears to have occurred rapidly and resulted in a shallow sea over the region. This early phase of sedimentation occurred on a stable platform in the north, but the more proximal southern parts of the basin were less stable and iron-formations and coeval volcanic rocks were deposited partly in grabens, that had been previously interpreted to be second order extensional features resulting from oblique collision of the continental margin with the Wisconsin magmatic terranes (Schneider et al., 2002). Alternatively, these grabens, which were contemporaneous with extension, bimodal calc-alkaline volcanism and massive sulfide deposition in the Pembine–Wausau arc terrane to the south, may have formed in response to back-arc extension after subduction flipped from south-directed to north-directed following the initial accretion of the Pembine–Wausau arc (see discussion below).

This initial phase of foreland sedimentation was followed by a significant hiatus of poorly understood tectonic significance. Recent geochronologic data indicate that the hiatus may have been basin-wide, in contrast to previous interpretations of continuous sedimentation in the northern parts of the basin. In Michigan, this hiatus has been long recognized as resulting in a low-angle unconformity. In the Gunflint and Mesabi Ranges, the physical record of the hiatus is not as obvious, but radiometric ages of rocks near the top of the Gunflint Iron-formation and the base of the Rove Formation indicate a period of roughly 40 m.y. of which only a thin sedimentary record is preserved. Recent detailed examination of this interval related to studies of the Sudbury impact layer provided new evidence that the 1850 Ma impact layer was deposited subaerially further documenting this somewhat cryptic sedimentary discontinuity (Addison et al., 2005). On the Mesabi Range, this interval has yielded no evidence of a hiatus and continuous deposition from the iron-formation to overlying clastic rocks has long-been assumed.

The major phase of foreland basin sedimentation consisted of a rapidly deepening basin into which a great thickness of turbidites was deposited. This began at about 1850 Ma in Michigan, based on identification of the Sudbury impact layer within the lowermost units of this sequence (Cannon et al., 2006). In northern parts of the basin sedimentation began significantly later, at about 1835 Ma, based on the age of volcanic ash layers near the base of the Rove Formation (Addison et al., 2005). Thus, there is clear evidence of northward migration of the foredeep over 10–15 m.y. Turbidite sedimentation in Michigan appears to have lasted no more than about 20 m.y. These rocks were incorporated into the foreland fold and thrust belt, deeply buried, and metamorphosed by 1830 Ma (Schneider et al., 2002). Metamorphic pressures indicate depths of burial as great as 15 km (Attoh and Klasner, 1989). Sedimentation appears to have been immediately north of and coincident with, the advancing arcs and foreland thrust sheets because both arc and reworked foreland detritus constitutes the bulk of the turbidites (Barovich et al., 1989). Post-tectonic intrusions at 1830–1835 Ma (Schneider et al., 2002) show that contractional deformation was essentially completed by that time. In the northern distal parts of the basin, sedimentation resumed significantly later than in Michigan and continued well past the end of Penokean deformation. The earliest beds were deposited at about 1835 Ma, at a time when lithostratigraphically equivalent sedimentary rocks in Michigan were already deeply buried and metamorphosed in the fold and thrust belt. Detrital zircons as young as 1780 Ma in upper parts of the Rove Formation indicate that deposition in the northern part of the basin outlasted Penokean deformation by as much as 50 m.y. (Heaman and Easton, 2005). This raises the question of the true age of the upper parts of the equivalent Virginia and Thomson Formations in Minnesota. We suggest the possibility that the continued subsidence of the foreland basin long after the end of Penokean overthrusting may reflect an as yet undocumented influence of Yavapai deformation and foreland subsidence. Perhaps the unconformable relationship between the Animikie Group and the rocks in the proximal foreland of Minnesota recorded a southward migration of the basin during the Yavapai orogeny.

2.2. *Tectonics of the continental foreland*

In this section we consider the tectonic evolution of the Penokean continental foreland that includes the Paleoproterozoic strata described above and their Archean basement. We define the tectonic foreland as the belt of folded and thrust-faulted rocks extending northward from the Niagara fault in Wisconsin and Michigan (Fig. 2). Roughly comparable terranes are known in Minnesota, but the southern boundary of the foreland is not as well defined because of both sparse outcrops and an abundance of younger post-Penokean granite intrusions, which have largely obliterated the southern part of the Penokean foreland (Fig. 2). The structure and tectonic evolution of this belt have been intensively studied and well mapped within the limits of available outcrops and the general structural setting and style are well established. This wealth of data is summarized in several modern papers (Klasner et al., 1991; Southwick and Morey, 1991; Holst, 1991).

2.2.1. *Michigan and Wisconsin*

In Michigan and Wisconsin two phases of deformation have been recognized within the Paleoproterozoic rocks (Cannon, 1973; Klasner, 1978) and both were traditionally ascribed to the Penokean orogeny. An earlier thin-skinned deformation resulted in northward-directed thrust sheets containing folded Paleoproterozoic sedimentary and volcanic rocks having a pervasive axial planar cleavage. There is little evidence that Archean basement rocks were affected during this phase of deformation, because pre-deformation diabase dikes in basement rocks generally were not deformed indicating that the Archean basement remained a rigid structural unit. A latter phase of doming and block faulting resulted in basement-cored uplifts having intervening structural keels, such as the Republic syncline. It is now well established that much of this basement uplift phase of deformation was substantially younger than the thin-skinned phase. Based on $^{40}Ar/^{39}Ar$ mica and hornblende ages and U–Pb monazite ages, the uplift event occurred at approximately 1760 Ma, about 70 m.y. after the emplacement of post-Penokean granite plutons (Schneider et al., 1996, 2004; Tinkham and Marshak, 2004; Tohver et al., 2007). Thus, the basement uplift phase cannot be considered a Penokean event. The cause of the basement uplifts remains unresolved. A passive uplift has been proposed in which crustal melting and gravitational collapse of the Penokean orogen led to rapid uplift and cooling of overthickened crust (Schneider et al., 1996, 2004). However, because the age of this uplift coincides with that of compressive deformation within the Yavapai orogen (Karlstrom et al., 2001), and because there is increasing evidence for Yavapai accreted terranes in central Wisconsin, only about 200 km south of the Penokean foreland (NICE Working Group, 2007), we suggest that an active tectonic mechanism related to Yavapai orogenesis can not be ruled out at this time.

In any case, the important point for this paper is that much of the structure of the Penokean foreland resulted from a post-Penokean event and that the effects of this younger event must be removed to reconstruct the Penokean foreland deformational history. We propose that Penokean foreland deformation was dominated by thin-skinned northward thrusting in which the Paleoproterozoic strata were structurally decoupled from Archean basement along detachment faults. Thus, the following descriptions are limited to that phase of deformation.

The foreland sedimentary sequence crops out for about 100 km north of the Niagara fault zone (Fig. 2) and was everywhere involved in the fold and thrust belt, although the intensity of the deformation decreases from the highly folded rocks in the south to the weakly deformed rocks in the north. The most northerly rocks, north of the Marquette Range, are for the most part gently dipping and were deformed into broad open folds, and commonly have well-developed slaty cleavage. Fold axes have near-horizontal plunges. Locally, the basal contact with the Archean basement rocks is exposed and is not a fault. Because stratigraphically higher units have been horizontally shortened as indicated by such features as deformed concretions, detachment faults can be inferred in the lower part of the sequence. Only a few faults of this type, however, can be seen in outcrop (Gregg, 1993). Farther south, in the central part of the Michigamme basin, a thick sequence of turbidites devoid of marker beds was folded on east trending axes. Beds generally dip steeply and extensive areas of uniformly facing structures, shown by graded beds, imply that fold wavelengths are as much as several kilometers. Still farther south, rocks within about 20 km of the Niagara fault were much more intensely and complexly deformed (LaBerge et al., 2003). Several thrust panels contain rocks that were multiply deformed and are characterized by mostly steeply plunging folds. The severe deformation clearly seems to have resulted because of the areas proximity to a major terrane boundary, the Niagara fault. The extraordinary deformation at that boundary resulted from the suturing of the foreland with the Wisconsin magmatic terranes during the culmination of the Penokean orogeny (Larue and Ueng, 1985; Ueng and Larue, 1987; Sedlock and Larue, 1985; LaBerge et al., 2003). The lack of Archean rocks in any of these panels, in spite of the widespread occurrence of the basal units of the Paleoproterozoic section, implies that the strata in the fault panels were detached from the Archean basement. As in areas to the north, even these most proximal parts of the fold and thrust belt experienced thin-skinned deformation and now constitute a multiple repetition of strata of the most proximal parts of the foreland depositional basin.

2.2.2. Minnesota

The Penokean deformation in Minnesota was broadly similar to that in Michigan and Wisconsin. An intensely and complexly deformed series of thrust panels on the south (Cuyuna North, Cuyuna South, Moose Lake, McGrath-Little Falls panels) gives way northward to progressively more weakly and simply deformed rocks across a belt about 100 km wide, succeeded farther north by essentially undeformed strata in the Mesabi and Gunflint Iron Ranges (Holst, 1991). Substantial progress has been made in deciphering the structure of the poorly exposed rocks of the Minnesota foreland through the use of aeromagnetic and gravity data and drillhole information. Southwick and Morey (1991) and Southwick et al. (1988) have presented syntheses of this information. The complex thrust panels on the south, like comparable structures in Michigan, appear to be thin-skinned slices without Archean basement (Fig. 2). However, as in Michigan, this area of thin-skinned thrusting is also the area in which Archean-cored gneiss domes developed during postorogenic collapse of the Penokean orogen (Holm and Lux, 1996; Schneider et al., 2004). Farther north, basement-cover relations are not well known other than in the Mesabi Range where Paleoproterozoic strata are mostly nearly flat lying above an undisturbed unconformity with Archean basement rocks.

3. The internal domain—Wisconsin magmatic terranes

Two distinct magmatic terranes are currently recognized, on the basis of rock type and structure, to constitute the Wisconsin magmatic terranes. The northern Pembine–Wausau terrane in Wisconsin is separated from the continental margin domain to the north by the Niagara fault zone, interpreted to be the southern fault in the Niagara suture zone (Fig. 2; LaBerge et al., 2003). Although the Niagara fault zone was previously extended westward to the Malmo discontinuity in eastern Minnesota (Southwick and Morey, 1991), more recent studies suggested that correlations of that structure with the Niagara fault zone are problematic (Schneider et al., 2004). It appears more likely that the Niagara fault trended southward in Minnesota and has been largely obliterated by the younger East-central Minnesota batholith and related granite plutons (Chandler et al., 2007). The southern Marshfield terrane is separated from the Pembine–Wausau terrane by the Eau Pleine shear zone, a major northwest-trending structure (Sims et al., 1989) that is also presumed to be a paleosuture (Fig. 2). The Marshfield terrane is truncated to the south in central Wisconsin by the northeast-trending Spirit Lake tectonic zone, a Yavapai-age structure perhaps correlative with the Cheyenne Belt (NICE Working Group, 2007).

3.1. Pembine–Wausau terrane

Rocks of the Pembine–Wausau terrane are exposed in northern Wisconsin over a strike length of 275 km and across a width of 150 km. However, it has been interpreted to extend in the subsurface to the west into East-central Minnesota (Southwick and Morey, 1991) and east to the extension of the Spirit Lake tectonic zone (NICE Working Group, 2007). The volcanic rocks in the terrane are mostly tholeiitic and calc-alkaline and have primitive oceanic arc to evolved island arc compositions. These rocks were deposited between about 1860 and 1889 Ma (Sims et al., 1989). Distinctly bimodal calc-alkaline basalt-low-SiO_2 andesite and dacite-rhyolite volcanic rocks locally host volcanogenic massive sulfide deposits (DeMatties, 1989; Sims et al., 1989; Schulz and Nicholson, 2000) and appear to have been deposited at about 1870 Ma (Sims et al., 1989; T. DeMatties, Geological Consultant, Minnesota, 1995, personal communication from R. Thorp, Geological Survey of Canada). Thus, these bimodal volcanic rocks and their contained massive sulfide deposits formed contemporaneously with the deposition of iron-formation and tholeiitic basalt in the Menominee Group in the foreland basin to the north. A more restricted dominantly calc-alkaline felsic volcanic succession was deposited between about 1835 and 1845 Ma on the older rocks along the southern margin of the terrane north of the Eau Pleine shear zone (LaBerge and Myers, 1984).

The older volcanic rocks in the Pembine–Wausau terrane generally have a prominent, steep east-trending foliation that is axial planar to tight folds. Metamorphism generally was in middle to upper greenschist facies, but locally reached amphibolite facies, particularly adjacent to gneiss domes such as the Dunbar dome in northeastern Wisconsin (Sims et al., 1985, 1992). The younger 1835–1845 Ma age volcanic rocks in the southern part of the terrane were folded about moderately open, northeast-trending, steeply plunging axes and were metamorphosed to middle to upper greenschist facies (LaBerge and Myers, 1984).

Granitoid rocks constitute nearly half of the outcropping rocks in the terrane and range in age from about 1889–1760 Ma (Sims et al., 1989). These intrusive rocks are mainly granodiorite and tonalite but include gabbro, diorite and granite (Sims et al., 1993). An older suite of dominantly calcic to calc-alkaline granitoids ranging in age from about 1889–1870 Ma appears to be cogenetic with the volcanic arc magmatism, while 1860–1840 Ma plutons are broadly contemporaneous with collision of the terrane with the Superior craton margin (Sims et al., 1992). Younger post-tectonic alkali-feldspar granite suites were emplaced at about 1835 and 1760 Ma (Sims et al., 1989).

Sims et al. (1989) interpreted the Pembine–Wausau terrane to have formed as an oceanic island arc with no significant continental basement in response to the southward subduction along the southern Superior craton margin, beginning by about 1900 Ma. Continued subduction led to the eventual collision of the arc complex with the continental margin by about 1860 Ma, the development of collision-zone intrusive bodies, such as those of the Dunbar dome (Sims et al., 1985, 1992), the formation of a foreland fold-thrust belt immediately north of the Niagara fault (suture) zone in Michigan (Klasner et al., 1988), and the northward translation of the continental margin depositional prism onto the craton (i.e., the Penokean orogeny). Subsequently, Van Wyck and Johnson (1997), on the basis of Pb and Nd isotope data and limited U–Pb zircon geochronology on granitoid rocks within the terrane, suggested a model in which the Pembine–Wausau terrane evolved as a continental arc-back-arc system on the southern margin of the Superior craton in response to northward subduction, which eventually led to collision with the Marshfield terrane to the south. Van Wyck and Johnson (1997) dated a strongly foliated tonalitic gneiss about 14 km south of the mapped trace of the Niagara fault zone in the northern part of the Pembine–Wausau terrane and obtained an Archean U–Pb zircon crystallization age of 2607 ± 22 Ma and a whole-rock Nd model age of 3.0 Ga. They also noted a general trend of lower ε_{Nd} values (i.e., greater crustal component) in plutons northward in the terrane and Pb isotopic compositions compatible with contamination by Archean crust of the Superior craton.

Particularly important to the tectonic interpretation of the Pembine–Wausau terrane is the presence of a dismembered suprasubduction zone ophiolite in northeastern Wisconsin along and south of the Niagara fault zone (Schulz, 1987; LaBerge et al., 2003; Schulz and Schneider, 2005). The Pembine ophiolite consists of tholeiitic basaltic and boninitic pillowed, massive, and fragmental volcanic rocks, massive to layered gabbros locally cut by sheeted mafic dikes, and ultramafic rocks (pyroxenities and serpentinites). The ophiolite is overlain by a sequence of calc-alkaline andesitic to rhyolitic lava flows and volcaniclastic rocks (Sims et al., 1989; LaBerge et al., 2003; Schulz and LaBerge, 2003). The tholeiitic basalts in the ophiolite sequence are compositionally similar to recent MORB and primitive oceanic island arc basalts whereas the boninitic rocks are similar in composition to Phanerozoic boninitic suites formed during the early stages of subduction of oceanic arcs (Shervais, 2001; Stern, 2002; LaBerge et al., 2003; Schulz and LaBerge, 2003). The volcanic rocks in the ophiolite have large positive ε_{Nd} values (~4.2; Beck and Murthy, 1991) supporting the interpretation that they were derived from depleted mantle with no continental crustal contamination. The overlying calc-alkaline volcanic rocks have somewhat lower (more enriched) positive ε_{Nd} values that are attributed to subduction of continental-derived sediments (Schulz and Ayuso, 1998). Attempts to obtain zircons to directly date rocks within the ophiolite have proven unsuccessful, probably because of the low zirconium content of these primitive arc rocks. However, a sill-like quartz diorite body, which intruded the upper part of the ophiolite sequence and is compositionally similar to the overlying calc-alkaline andesites, has a U–Pb zircon age of 1889 ± 6 Ma (Schulz and Schneider, 2005), providing a minimum age for the ophiolite sequence. The presence of this suprasubduction zone ophiolite-calc-alkaline arc sequence along the northern margin of the Pembine–Wausau terrane, that has characteristics compatible to the birth to maturity stages of evolution of an oceanic arc (Shervais, 2001; Schulz and LaBerge, 2003), strongly suggests that the terrane formed as a Paleoproterozoic oceanic island arc in response to southward (present coordinates) directed subduction (Schulz and Schneider, 2005). This would not preclude the possible presence of older, perhaps Archean, basement blocks locally within the terrane. It does suggest, however, that such older basement was not present throughout the terrane, at least prior to collision with the Superior craton margin.

Schulz and Ayuso (1998) analyzed Nd and Pb isotopes in volcanic rocks and selected granitoids from the Pembine–Wausau terrane to further investigate the contribution of older basement to the magmatic evolution of the terrane. They documented that the volcanic rocks across the terrane generally have higher positive ε_{Nd} values than the associated intrusive granitoids, particularly some of the collisional and post-tectonic plutons that were intruded from about 1860–1835 Ma. For example, in contrast to the Pembine ophiolite and associated calc-alkaline volcanic and intrusive rocks, which have positive ε_{Nd} values, samples of Dunbar Gneiss, which was emplaced into the Pembine ophiolite-calc-alkaline volcanic arc sequence at 1862 ± 5 Ma (Sims et al., 1985), have negative ε_{Nd} values and whole-rock Nd model ages ranging from 2.29 to 2.41 Ga indicating a significant component of continental crust. In addition, the post-tectonic Bush Lake granite (~1835 Ma), that intruded rocks on the northwest side of the Dunbar dome (Sims et al., 1985, 1992), has a large negative ε_{Nd} value and an Archean Nd model age (Schulz and Ayuso, 1998). These data clearly suggest a change in source region from depleted mantle during formation of the ophiolite and associated calc-alkaline arc to dominantly continental (Archean) crust during formation of the younger collisional to post-tectonic granitoid intrusions.

We suggest that this change in source region resulted from the obduction and overthrusting of the Pembine–Wausau arc terrane onto the southern margin of the Superior craton during the Penokean orogeny. Obduction of oceanic arcs onto continental margins is a typical event when continental margins are subducted beneath colliding arcs (Shervais, 2001) as witnessed today by relations in Taiwan, Timor and Papua New Guinea in the southwest Pacific (Huang et al., 2000). This interpretation is supported by the gravity modeling of Klasner et al. (1985) and Attoh and Klasner (1989), who inferred that Archean crust, defined by a broad, long-wavelength, positive gravity anomaly, extends at least 30 km in a southwesterly direction across the Niagara fault from northern Michigan into Wisconsin to underlie the northern and western parts of the Pembine–Wausau terrane. Further, two-dimensional modeling by Attoh and Klasner (1989) of a north-south gravity profile across the Dunbar dome northward into northern Michigan suggested that the Pembine–Wausau terrane is allochthonous and was transported northward onto the continental foreland. Thus, the strongly foliated Archean tonalite dated by Van Wyck and Johnson (1997) south of the Niagara fault zone may be underthrust Superior crust exposed in an uplifted erosional window through the allochthonous arc terrane.

3.2. Marshfield terrane

The Marshfield terrane, south of the Eau Pleine shear zone (Fig. 2), has Archean gneisses exposed over more than 50 percent of its area (Sims, 1989). The gneisses are presumed to stratigraphically underlie the Paleoproterozoic volcanic rocks throughout all or most of its area. Because of a strong structural overprint and the generally sparse outcrops, knowledge of the Proterozoic rocks is still meager.

Paleoproterozoic volcanic rocks in the Marshfield terrane were deposited on Archean basement at about 1870–1860 Ma and are preserved only as erosional remnants (Sims, 1989). They consist principally of an interlayered sequence of felsic to mafic volcanic rocks, dacite porphyry, and a variety of sedimentary rocks including impure quartzite, ferruginous chert, conglomerate, and carbonaceous argillite (Sims et al., 1989). The conglomerate locally contains clasts of Archean granitic gneiss (Myers, 1980; Myers et al., 1980). The interlayered volcanic-sedimentary sequence was intruded by gabbro, diorite, and tonalite locally dated at about 1865 Ma (Van Wyck, 1995). The main area of volcanic and sedimentary rocks in the Marshfield terrane consists of a steeply plunging synform that is closed to the east; foliation generally dips steeply (>70°). The original structure has been largely obliterated by superimposed steep stretching elements that are nearly ubiquitous in both the Paleoproterozoic supracrustal rocks and the subjacent Archean basement. Metamorphism of the supracrustal rocks was mainly upper greenschist (garnet) grade and was concomitant with the superimposed ductile deformation. It has distinctly lower grade than the lower to middle amphibolite facies metamorphism that characterizes the crystalline rocks elsewhere in the terrane (Maass et al., 1980). The difference in metamorphism may reflect different crustal levels across the terrane.

The Paleoproterozoic tonalitic intrusive rocks in the Marshfield terrane range from hornblende-biotite tonalite to biotite trondhjemite and range in age from 1892 ± 9 to 1841 ± 24 Ma (Sims et al., 1989; Van Wyck, 1995). Also present locally are post-tectonic 1835 Ma alkali-feldspar granites, at places with lesser rhyolite, and 1760 Ma alkali-feldspar granites (Sims et al., 1989). The older tonalitic intrusive rocks are calc-alkaline in composition and have steep, heavy rare-earth element depleted patterns and small or no Eu anomalies (Anderson and Cullers, 1987; Van Wyck, 1995). Their chemistry is compatible with melting of an eclogitized basaltic source with residual garnet and/or hornblende (Anderson and Cullers, 1987; Van Wyck, 1995), although they could also have resulted from partial melting of Archean basement rocks which are themselves typically characterized by steep, heavy rare-earth element depleted patterns (Martin, 1994). Van Wyck and Johnson (1997) have shown that the Paleoproterozoic tonalities in the Marshfield terrane generally have lower ε_{Nd} values (-1 to -7) than do plutons in the Pembine–Wausau terrane to the north, suggesting a greater contribution of Archean crustal Nd to the plutons in the Marshfield terrane. Van Wyck and Johnson (1997) also noted that the Archean basement rocks and intrusive Paleoproterozoic plutons in the Marshfield terrane are characterized by less radiogenic Pb isotopes than those in the Pembine–Wausau terrane and Superior Archean crust. This supports the interpretation that the two terranes evolved separately and that the Archean basement in the Marshfield terrane may not be a rifted fragment from the Superior craton.

Sims et al. (1989) suggested that northward directed subduction led to the eventual collision and suturing of the Marshfield terrane to the Pembine–Wausau terrane. The time of ductile deformation within the Marshfield terrane and the joining of the Marshfield terrane to the Pembine–Wausau terrane along the Eau Pleine shear zone is constrained between about 1860 and 1835 Ma. Undeformed alkali-feldspar granites that intrude the Marshfield terrane, the Eau Pleine shear zone, and the southern margin of the Pembine–Wausau terrane range in age from 1853 ± 21 to 1833 ± 4 Ma (Sims et al., 1989) and have strongly negative ε_{Nd} values and mostly Archean Nd model ages (Schulz and Ayuso, 1998). We suggest that the northward directed subduction that led to the eventual collision of the Marshfield terrane with the Pembine–Wausau terrane was initiated after the start of collision of the Pembine–Wausau terrane with the Superior continental margin at about 1880 Ma. The collision of the Pembine–Wausau terrane would likely have terminated the southward subduction because the buoyancy of the continental margin crust would stop subduction and lead to a flip of subduction polarity as has been commonly observed in younger arc-continental margin collisions (Van Staal et al., 1998). The resulting north-directed subduction may have produced a period of back-arc extension in both the Pembine–Wausau terrane and the Superior continental margin in a manner similar to that proposed by Hyndman et al. (2005) for the western North America Cordillera region. This back-arc extension, active by about 1875–1870 Ma, may account for the second order grabens and tholeiitic volcanism in the evolving foreland basin on the continental margin and the bimodal calc-alkaline volcanism and massive sulfide deposition in the Pembine–Wausau terrane. Northward subduction also would have likely led to development of a marginal magmatic arc represented by the 1845–1835 Ma calc-alkaline volcanic rock north of the Eau Pleine shear zone (Sims et al., 1989). The collision of the Marshfield terrane with the Pembine–Wausau terrane at about 1850 Ma is suggested to have resulted in the major period of Penokean foreland deformation and sedimentation.

3.3. *Synopsis of magmatic arc development and accretion*

The Wisconsin magmatic terranes record a complex series of tectonic events related to Paleoproterozoic subduction and collision along the southern margin of the Superior craton. Initial southward subduction led to the development of an oceanic suprasubduction zone ophiolite and island arc complex that constitutes the Pembine–Wausau terrane. Subduction began some time before 1889 Ma and ended by about 1880 Ma (Fig. 4) with the relatively soft collision of the arc complex with the Superior continental margin. The obduction of the arc terrane onto the Superior margin probably resulted from the partial subduction of the Superior margin beneath the arc complex during initial collision. However, the obduction of the arc terrane appears to have principally resulted in downwarping of the continental margin and initiation of a foreland basin. The relatively "soft" collision of the Pembine–Wausau terrane with the continental margin may have been a result of the collision occurring in an embayment. Collision of a linear arc with an irregular continental margin will occur first along promontories and result in a decrease in the magnitude of convergence within the embayments (Stockmal et al., 1987). The volcanogenic massive sulfide deposits interbedded with bimodal calc-alkaline volcanic rocks at about 1870 Ma suggests a period of intra-arc extension (rifting) overlapping the time of volcanism in the developing continental foreland basin to the north (Schneider et al., 2002). This period of intra-arc rifting may have developed in response to northward subduction and back-arc extension prior to the collision of the Marshfield terrane. The Niagara fault zone now marks the frontal thrust along which arc and continental margin rocks were complexly interleaved (Sims et al., 1992; LaBerge et al., 2003). Suturing was completed by 1830–1835 Ma when undeformed, crustally derived, alkali-feldspar granites were intruded on both sides of the Niagara fault zone forming stitching plutons.

The Marshfield terrane records the evolution of a calc-alkaline arc on Archean basement initially some distance south of the Pembine–Wausau terrane between about 1890–1840 Ma. Following northward subduction, the two magmatic terranes were sutured along the Eau Pleine shear zone before 1835 Ma when undeformed, crustally derived, alkali-feldspar granites, locally with associated rhyolites, where emplaced within both terranes. The period of northward subduction led to back-arc extension in both the Pembine–Wausau terrane and the continental margin. The collision of the Marshfield terrane appears to have been a "harder" collision than that of the Pembine–Wausau terrane. It resulted in the uplift and erosion of the Pembine–Wausau terrane, and the rapid development of a fold and thrust belt, a migrating foreland bulge, and a turbidite depositional basin in the external continental margin domain. This harder collision may have resulted from the embayment having already been filed and, thus, the margin at least partially straightened by the prior accretion of the Pembine–Wausau terrane. It also may have been aided by the presence of Archean crust in the Marshfield terrane providing for collision of stronger and more rigid crust (Chandler et al., 2007).

4. Post-Penokean overprints

During the past decade refinements in the ages of various rocks units and tectonic events make it abundantly clear that rocks of the Penokean orogen, throughout much of its extent, carries a strong overprint of younger events. In keeping with our definition of the Penokean orogeny as being restricted to events older than the ca 1830 Ma post-tectonic plutons that cut Penokean structures, these events range from lingering thermal after-effects of the Penokean, which are only slightly younger than 1830 Ma, to pronounced and pervasive tectonic overprints that are as much as two geons antecedent to the close of the Penokean. Our purpose here is not to describe these features in detail, but rather to point out their existence and the necessity to see through these overprints in order to correctly ascribe various structures to their correct orogenic setting. In fact, this task is far from complete and significant reevaluation of the tectonic affinities of many features is still needed.

The first major overprint occurred in the interval 1775–1750 Ma. A region-wide period of copious granitic magmatism resulted in a suite of plutons, most notably the East-central Minnesota batholith, that trends along the southern foreland and northern internal belt of the Penokean orogen (Fig. 2), the region that probably experienced the greatest tectonic thickening during Penokean collision. Regional metamorphism to upper amphibolite facies accompanied the magmatism (Holm and Lux, 1996; Holm et al., 1998a, 2005). Also approximately coincident with the magmatism, both temporally and spatially, was development of Archean basement-cored gneiss domes and intervening troughs containing folded Paleoproterozoic strata (Holm and Lux, 1996; Schneider et al., 2004). The resulting gneiss dome corridor extends from East-central Minnesota across much of northern Michigan (Schneider et al., 2004). The regional map pattern of uplifted Archean rocks surrounded by Paleoproterozoic rocks, such as is evident on Fig. 2, is a result of this mid-geon 17 tectonism. Although the gneiss dome corridor follows the trend of the older Penokean orogen and has been traditionally interpreted to be a Penokean tectonic feature, it is critical to understanding the nature of the Penokean orogen that these structures be excluded from the inventory of truly Penokean features.

A still younger tectonic event appears to have affected most of Wisconsin and parts of northern Michigan. Quartzites of the “Baraboo interval” are widespread in Wisconsin and Minnesota. They all appear to be younger than 1750 Ma, the age of the youngest detrital zircons that they contain (Holm et al., 1998b), yet most of these rocks in Wisconsin are strongly folded. The quartzites in Wisconsin lie unconformably on rocks within the internal zone of the Penokean orogen (Fig. 2). The basement rocks below these deformed quartzites have cooling ages of 1630 Ma. The folding is interpreted to have occurred during the foreland deformation accompanying the Mazatzal orogeny, which was active south of this region in mid-geon 16 time (Holm et al., 1998b). Although the folding of the quartzites during the Mazatzal orogeny seems unequivocally established, it has not been clearly distinguished from the Penokean orogeny in the underlying volcanic rocks. Because Mazatzal and Penokean folding was approximately coaxial, such differentiation of folding events is difficult on purely geometric relations. At present, the tight folds within rocks in the volcanic belts of the Wisconsin portion of the internal zone cannot be used to clearly separate folds formed during the Penokean orogeny from those formed during the later Mazatzal orogeny. It is quite possible that some folds previously ascribed to Penokean deformation are, in fact, Mazatzal structures.

5. Synopsis of the history of the Penokean orogen

A set of schematic cross-sections illustrating the tectonic evolution of the Penokean orogen is shown in Fig. 6. The scheme shown and expanded on in the following discussion is compatible with most of the current knowledge of the orogen as presented above. Our interpretation, however, is not unique and variations have been proposed by others in recent years. Fig. 6a shows the proposed arrangement of tectonic elements at about 1890 Ma. By this time, ocean closure had begun and the Pembine ophiolite and succeeding calc-alkaline arc, now constituting the Pembine–Wausau terrane, had largely formed and had been intruded by synvolcanic batholiths, both having ages as old as 1890 Ma. By 1875 Ma (Fig. 6b) a “soft” collision of the arc with the embayment along the southern edge of the Superior craton was well underway. Sediments of the associated basin, the Menominee Group in Michigan and the North Range and lower part of the Animikie Group in Minnesota, were being deposited as indicated by recent radiometric ages of ca. 1875 Ma for the volcanic rocks interlayered with the iron-formations. This implies that collision of the arc terrane with the southern edge of the Superior craton was underway by about 1880 Ma and that tectonic loading of the craton margin had formed a broad, shallow, basin in which northerly derived siliciclastic sediments had been deposited on Archean basement. These rocks were succeeded by major iron-formations and lesser coeval bimodal tholeiitic basalt and rhyolite, locally in second order extensional grabens. Because the Superior craton was impinging on the Niagara subduction zone at 1880 Ma, it is likely that the physical restrictions on subducting thick continental lithosphere inhibited further southward subduction and resulted in a southward jump in subduction to a position south of the Pembine–Wausau arc and a flip of subduction polarity to the north. The northward subduction led to a period of back-arc extension during which bimodal calc-alkaline volcanic rocks and associated massive sulfide deposits formed in the Pembine–Wausau arc terrane and second order extensional grabens with associated bimodal tholeiitc volcanism formed in the continental margin. The presence of comparable age calc-alkaline volcanic rocks and related intrusions in the Marshfield terrane suggests that it was also part of an evolving arc system at this time. Note that the original extent of the Marshfield terrane is not known (Fig. 6b). Recent interpretation of regional geophysical data suggests that the Marshfield terrane is truncated by a younger, probably Yavapai age, structure, the Spirit Lake tectonic zone (NICE Working Group, 2007).

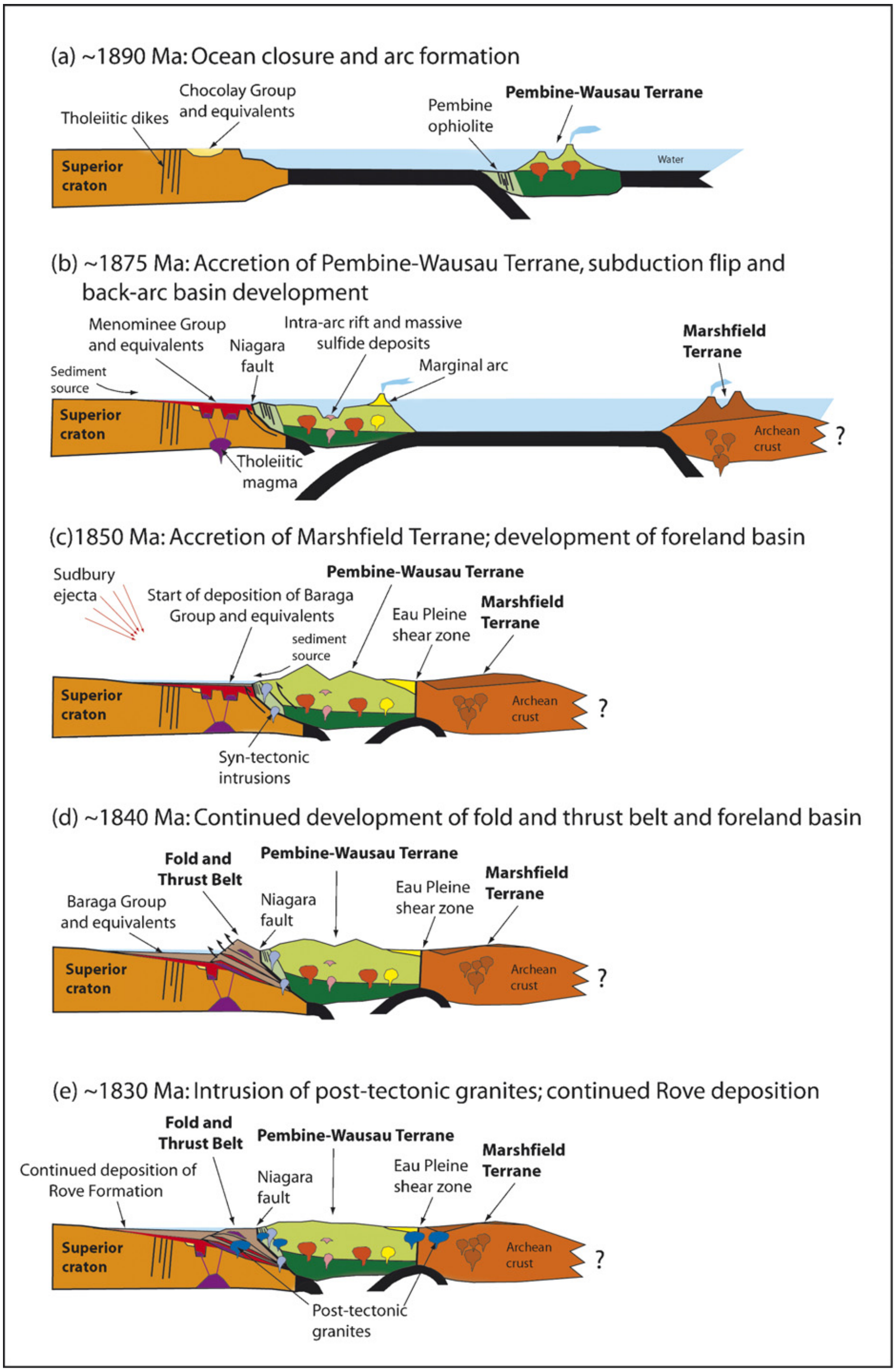

Fig. 6. Schematic cross-sections illustrating the tectonic evolution of the Penokean orogen. See text for discussion.

By 1850 Ma (Fig. 6c), subduction had stopped as the Marshfield terrane collided with the Pembine–Wausau terrane. Initial collision of two terranes resulted in a fore-bulge that migrated across the developing foreland basin as the Pembine–Wausau terrane was uplifted and thrust craton-ward. The deformation of the Pembine–Wausau terrane was accompanied by emplacement of syn-tectonic intrusions south of the Niagara fault zone. Onset of sedimentation in the foreland basin was marked by the deposition of ejecta from the Sudbury impact.

Fig. 6d shows the inferred conditions at 1840 Ma, near the close of the Penokean orogeny. Compression along the Eau Pleine shear zone had led to continued compression of the Pembine–Wausau terrane and tectonic loading of the margin of the Superior craton with thrust sheets of volcanic and sedimentary rocks forming a fold and thrust belt. The increasing load resulted in a deep foreland basin into which a great thickness of southerly derived turbidite deposits accumulated. Turbidite sedimentation initiated at roughly 1850 Ma. Sediments near the southern edge of the basin were buried as deep as 15 km by sedimentation and nearly coeval overthrusting. These rocks were metamorphosed to upper amphibolite facies by 1830 Ma when deformation had ended and post-tectonic granites were intruded across the orogen (Fig. 6e).

The geology and tectonic history of the Penokean orogen summarized above are similar to those of more recent orogens (e.g., Van Staal et al., 1998) and strongly suggest that plate tectonic processes were well established in the Paleoproterozoic. Examination of recent cognate arc systems has shown that, although they can vary greatly in length, typically they have lengths of more than 1000 km (Van Staal et al., 1998). However, recent data regarding the lateral extent of the Penokean orogen indicate that Penokean aged rocks and deformation are largely confined to the Lake Superior region in the area of the former embayment of the Superior craton (NICE Working Group, 2007). This implies that post-Penokean tectonic events may have removed or overprinted significant parts of a once more extensive Penokean orogen.

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