LATE- TO POST-TECTONIC SETTING OF SOME MAJOR PROTEROZOIC
ANORTHOSITE – MANGERITE – CHARNOCKITE – GRANITE (AMCG) SUITES

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ABSTRACT

Recent zircon-based geochronological investigations in the Adirondack Mountains, in the state of New York, demonstrate that 1155 ± 6 Ma massif anorthosite was emplaced during the post-contractional phase (1160–1140 Ma) of the ca. 1200–1140 Ma Shawinigan orogeny. Emplacement of many other anorthosite – mangerite – charnockite – granite (AMCG) suites also correlates with the waning stage of orogeny and includes the ca. 1155 Ma Morin and Lac-St-Jean complexes, the ca. 1650 Ma Mealy Mountain complex, and the ca. 1450 Ma complexes of eastern Labrador. The correlation also applies to the ca. 930 Ma Rogaland anorthosite complex of Norway and the ca. 1060–1020 Ma late- to post-Ottawan anorthosites of central Quebec and the Appalachians of Virginia and southeastern Pennsylvania. These correlations suggest models involving delamination of overthickened orogenic lithosphere by foundering or convective removal, followed by asthenospheric ascent, and ponding of gabbroic melt at the crust–mantle interface. Orogen rebound following delamination results in stable, dynamically balanced settings in which gabbroic magma evolves slowly at high pressure to produce high-aluminum pyroxene and coarse, intermediate plagioclase characteristic of massif anorthosite. Related melting of the lower crust produces mangeritic and charnockitic magma. Ultimately, both anorthositic and granitic magmas ascend, and lower-pressure fractionation yields the classic AMCG suites. Transtensional reactivation of lithospheric-scale shear zones and old sutures also correlates with important AMCG magmatism and provides conduits for gabbroic magma that ponds at the crust–mantle interface or in the deep crust to produce AMCG suites. Flat-slab subduction, back-arc extension, slab breakoff, and hotspots represent alternative settings that can account for gabbroic underplating and fractionation into AMCG suites, if consistent with geochronological constraints.

Keywords: anorthosite, post-orogenic, geochronology, delamination, shear zone, AMCG (anorthosite – mangerite – charnockite – granite) suites.

Sommaire

Nos travaux géochronologiques récents portant sur les monts Adirondack, état de New York, démontrent que les massifs d’anorthosite ont été mis en place à 1155 ± 6 Ma, lors de la phase post-contractionnelle (1160–1140 Ma) de l’orogénèse Shawinigan, dans l’intervalle ca. 1200–1140 Ma. La mise en place de plusieurs autres suites de type AMCG (anorthosite – mangerite – charnockite – granite) semble associée aux stades finaux d’une orogénèse, par exemple ca. 1155 Ma pour les complexes de Morin et de Lac-St-Jean, ca. 1650 Ma pour le complexe de Mealy Mountain, ca. 1450 Ma pour les complexes de l’est du Labrador. La corrélation s’applique aussi au complexe anorthositique de Rogaland, en Norvège, à ca. 930 Ma, et les anorthosites du centre du Québec et des Appalaches en Virginie et de Pennsylvanie, liées au stades tardifs de l’orogénèse Ottawa à ca. 1060–1020 Ma. Ces corrélations font penser à un événement de délamination par affaissement d’une lithosphère orogénique surépaisse ou bien par érosion convective, ce qui aurait provoqué une montée de manteau asthénosphérique, et accumulation de magma gabbroïque à l’interface croûte–manteau. Le soulèvement de l’orogène suivant la délamination a produit un contexte stable, dynamiquement équilibré, dans lequel le magma gabbroïque a évolué lentement à pression élevée pour produire un clinopyroxène alumineux et un plagioclase intermédiaire à grains grossiers, caractéristiques de l’anorthosite en massif. En même temps, unefusion de la croûte inférieure a produit les magmas mangeritiques et charnockitiques. Éventuellement, les magmas

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anorthositiques et granitiques migrent vers la surface, et leur fractionnement à plus faible pression produit l’association classique des suites AMCG. Une réactivation transtensionnelle le long de zones de cisaillement à l’échelle lithosphérique et d’anciennes sutures montre aussi une corrélation avec le magmatisme AMCG, et fournit des conduits pour le magma gabbroïque, qui s’accumule à l’interface croûte–manteau ou bien dans la croûte profonde pour produire les associations AMCG. Une subduction subhorizontale, une extension derrière l’arc, l’affaissement d’une plaque, et l’existence de panaches mantelliques représenteraient d’autres contextes possibles pour expliquer la venue de magmas gabbroïques et leur fractionnement pour donner les suites AMCG, compte tenu des contraintes géochronologiques.

(Traduit par la Rédaction)

**Mots-clés:** anorthosite, post-orogénique, géochronologie, délamination, zones de cisaillement, suites AMCG (anorthosite – mangerite – charnockite – granite).

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**INTRODUCTION**

Proterozoic massif anorthosites have attracted interest for many decades owing, in part, to their large size, almost monomineralic compositions, occurrence as plutonic rocks only, and their somewhat restricted occurrence (Fig. 1) to Proterozoic belts (Herz 1969, Rämö *et al.* 2003). They are characterized by composite batholiths consisting of multiple, coalesced plutons of variable composition, but dominated by anorthosite and leuconorite. Although many of their petrological problems have been resolved, the issue of their tectonic setting is still unsettled, and their spatial–temporal restriction remains enigmatic. Ashwal (1993), Emslie (1978, 1985), Emslie *et al.* (1994), McLelland *et al.* (1996, 2004), Corrigan & Hanmer (1997), Scoates (1994, 2008), and Morisset *et al.* (2009), among others, have summarized these issues.

In the present paper, we provide evidence and arguments that stress 1) the importance of gabbroic parental magmas that pond and differentiate under relatively quiescent conditions at the crust–mantle interface, yielding anorthositic derivatives that ascend into the middle to upper crust together with granitoids derived by crustal anatexis, and 2) results of precise zircon geochronology that places these events in late to post-tectonic settings. Such settings are prone to delamination of overthickened lithosphere and to post-collisional extension promoting the influx and ponding of gabbroic magma formed by decompression melting.

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**Fig. 1.** Distribution of AMCG suites and ferroan-potassic (A-type and rapakivi) granites across North America and western Europe. Symbols: AD: Adirondack Mountains, G: Grenville orogen, HC–L: Horse Creek and Laramie AMCG complexes, M: Mazatzal orogen, N: Naïn AMCG complex, R: Rogaland AMCG complex in the Sveconorwegian orogen, SF: Saint Francois Mountains (black cross), SG: San Gabriel AMCG complex, Y: Yavapai orogen. Dashed lines approximate boundaries of orogenic belts. Modified after Haapala & Rämö (1999).
Such magmas can also be tapped by the interaction of steep shear-zones with ancient crustal boundaries. Examples of anorthosites in all of these settings are presented along with relevant zircon geochronology.

**PETROLOGICAL CONSTRAINTS ON ANORTHOSITE GENESIS**

There has been general agreement (see Ashwal 1993 for a summary) that massif anorthosite is derived from a high-aluma gabbroic magma ponded at the crust–mantle boundary, where it underwent early (Phase 1) fractionation of olivine and pyroxene that sank back into the mantle, as well as fractionation of intermediate plagioclase that floated in the increasingly dense residual magmas (Kushirow 1980, Scoates 2000). Ultimately, plagioclase-rich crystal mushes ascended into the middle and upper crust, where the stability field of plagioclase expanded and liquid fractionation undergone lower-pressure differentiation (Phase 2) that yielded more anorthosite and leucogranite as well as small volumes of late ferrodiorite and Fe–Ti oxide ores. During early ponding, advected heat, as well as heat of crystallization of the high-aluma gabbros, caused partial melting in the lower continental crust, resulting in quartz-poor, potassic, iron-rich, orthopyroxene-bearing granitoids that, together with the anorthositic magmas, form coeval (but not comagmatic) anorthosite – mangerite – charnockite – granite (AMCG) suites. Experimental data (Wyllie 1977) indicate that high-pressure, anhydrous melting of continental crust produces silica-poor, potassium-rich minimum melts that will fractionate during ascent into orthopyroxene-bearing A-type granitoids similar to mangerites, charnockites, and granites. In addition, fractionation of magnesian mafic phases will drive the compositions toward iron enrichment characteristic of the MCG rocks. The volume of MCG granitoids greatly exceeds that of anorthosite in most complexes and is consistent with a coeval, but not comagmatic origin. In a competing petrogenetic model proposed by Duchesne, Vander Auwera and coworkers (e.g., Duchesne et al. 1999, Vander Auwera et al. 1998), mantle–crust interaction during partial melting of slabs of anhydrous gabbroic lower crust thrust into the upper mantle produces jotunitic magmas that fractionated plagioclase (i.e., anorthosite) and evolved toward silica enrichment (i.e., the MCG granitoids). Scoates (1994, 2008) has shown that these jotunitic magmas cannot produce olivine-normative magmas characteristic of the early phases of many anorthosites, nor do they follow the iron-enrichment, silica-depletion differentiation trend characteristic of anorthositic magmas (Scoates 1994, 2008, Morisset et al. 2009).

It is generally agreed that a link exists between layered mafic intrusions and massif anorthosites (Morse 1968, Ashwal 1993), and it is not uncommon to find the two spatially associated (e.g., Kiglapait and Nain complexes, Morse 1968). The lineage is to be expected, since both are products of gabbroic magmatism, but identifying and explaining the causes of the differences between the two are other matters. We propose that to a large extent, the differences reflect variations in both depths of fractionation and magma-residence time at those depths, with anorthosites evolving at greater depths and for longer times than mafic layered intrusions. In what follows, we shall endeavor to demonstrate how the appropriate conditions for the generation of massif anorthosite arise.

Emslie (1975) noted that ponding of a high-aluma gabbroic magma at the mantle–crust interface is consistent with the presence of megacrysts (5–50 cm and up to 1 m) of aluminous pyroxene (most commonly orthopyroxene) within many massif anorthosites. These may occur as subrounded single crystals or groups of crystals (Type I) or in subophitic intergrowth with equally coarse plagioclase (Type II), the latter type forming rafts of megacrystic anorthosite within finer-grained anorthosite and leucogranite. Both types of megacrysts have elevated concentrations of Al, but Type 2 is especially high, ranging from ~4 to 9 wt.% Al₂O₃, indicative of pressures of crystallization of 5–12 kbar or ~15–40 km depth (Green 1969, Fram & Longhi 1992, Longhi et al. 1999). The high-pressure crystallization documented by the pyroxene megacrysts is consistent with compositions of associated megacrysts of plagioclase (An₄₈–₅₅) in the coarse rafts. As shown experimentally by Fram & Longhi (1992), the An content of plagioclase in equilibrium with high-aluma gabbro decreases with rising pressure so that the composition of plagioclase on the liquidus changes from An₇₄ at 1 kbar to An₃₄ at 11.5 kbar, which overlaps megacrystic plagioclase (~An₄₈–₆₀) in the massifs and is consistent with crystallization from a basic melt at pressures of 10–12 kbar. These pressures agree with those associated with giant megacrysts of high-Al pyroxene as well as the field evidence that both crystallized under the same deep conditions and were transported as rafts or crystal mushes to their present positions.

Also implied for the deep (Phase 1) environment are high-temperature, anhydrous conditions, and sufficient tectonic stability to permit the growth of large crystals and retention of melts for long periods of time. The absence of significant zoning within most plagioclase implies replenishments of new high-aluminum basic melt to keep magma composition approximately constant as well as to increase the quantity of plagioclase cumulate beyond that produced by a single melt. As described by Emslie et al. (1994) and Bickford et al. (2010), AFC interaction of the gabbros with pyroxene granulite restites in the lower crust imparts a mild, but very significant, crustal trace-element signature to them, and to their derivative anorthosites, and also extends the lifetime of plagioclase on the liquidus. In a related manner, Duchesne et al. (1999) argued that elevated values of ¹⁸⁷Os/¹⁸⁸Os in the Suwalki anorth-
site in Poland require a predominantly crustal source for the anorthosite. However, Hannah & Stein (2002) have shown that the Os isotope characteristics can be explained by assimilation of crustal materials by mantle-derived anorthositic magmas.

**FIELD RELATIONSHIPS AND TECTONIC SETTINGS OF ANORTHOSITE GENESIS**

The field relations of massif anorthosites have been clearly summarized by Emslie (1978), who pointed out that whereas these bodies commonly occur in high-grade terranes, they rarely, if ever, show evidence of being synchronous with regional contraction. Some are overprinted by later metamorphism (e.g., Marcy), and others appear to be pristine (e.g., Nain, Fig. 2). On the basis of such observations, anorthosites and their associated MCG granitoids came to be regarded as “anorogenic” plutonic rocks that, together with ferroan, potassic granitoids (i.e., A-type and rapakivi granites), formed a Proterozoic magmatic belt across North America (Fig. 1). Although the term “anorogenic” was never clearly defined, it has generally been understood to imply a region of mild to abortive rifting in which ponded basic magmas could evolve as described above (Emslie 1978, 1985). Within this environment, it was presumed that rifting was sufficiently slow to ensure that the ponded magmas were not prematurely removed from their site of fractionation, and that the absence of major contraction precluded large-scale mixing with crustal materials that would produce intermediate magmas. In short, an “anorogenic” environment was one of thick, tectonically stable crust that could provide the hot, high-pressure, tectonically stable, and anhydrous conditions required by petrological models keyed to deep, protracted fractionation of the basic magmas. In some cases, field evidence is consistent with a back-arc setting (e.g., Mistastin, Michikamau, and Harp Lake massifs: Rivers & Corrigan 2000, Rivers 2008b). In other instances, reactivation along major fault zones or terrane boundaries (e.g., the Horse Creek anorthosite: Scoates & Chamberlain 1995, 1997, Frost et al. 2000; the Nain complex: Myers et al. 2008) corresponds with field relations. Some investigators (e.g., Hoffman 1989) suggested that subcontinental hot spots and plumes could produce the conditions necessary for “anorogenic” magmatism, but demonstrating that this has proven to be elusive. All of these mechanisms are potentially consistent with the petrogenetic constraints outlined above; however, a critical boundary-condition for actual anorthosite genesis is provided by geochronology, which we discuss below.

**U–Pb ZIRCON GEochRONOLOGY OF AMCG SUITES**

By the 1980s, U–Pb zircon dating became widely accessible and was applied to high-grade terranes, including many in the Grenville Province (Fig. 2). Within the Adirondacks, McLelland et al. (1988b) produced ages by multigrain thermal-ionization mass spectrometry (TIMS) on an array of metamorphosed igneous rocks, including seven granitoid samples from the Marcy AMCG suite that ranged from 1155 to 1125 Ma (Table 1). Similar efforts by Emslie & Hunt (1990) in the central Grenville Province yielded five multigrain TIMS ages (1160–1123 Ma) for massif MCG suites, [excluding the Labrieville (1018 Ma), Rivière Pentécôte (1365–1350 Ma), and Mealy Mountains (1646–1635 Ma), all of which suggested more than one period of AMCG magmatism]. It was assumed on the basis of field relationships that all members of these suites were coeval, but in the absence of direct dating of anorthosite with zircon, this represented little more than a well-informed assumption.

Difficulty in dating anorthosite arose from the paucity of igneous zircon in these rocks. Silver (1969), in his landmark application of zircon geochronology to the Adirondacks, noted that the anorthosites contain numerous small, multifaceted grains of zircon having a “soccerball” morphology that yielded multigrain TIMS ages of ca. 1050 Ma, but he stressed that these zircon grains were clearly metamorphic in origin. Machado & Martignole (1988) and Martignole et al. (1993) obtained the first U–Pb zircon age for magmatic zircon from an anorthosite (i.e., 1354 ± 3 Ma for the Rivière Pentécôte intrusion in Quebec). Subsequent dating of charnockite associated with that massif yielded an age of ca. 1350 (Emslie & Hunt 1990), helping to support the coeval nature of AMCG suites. In Wyoming, Frost et al. (1993) obtained a single-grain zircon age of 1439 ± 7 Ma for a monzonitic rock from the Laramie anorthosite complex. Direct dating of the Adirondack anorthosite did not get under way until the J.C. Roddick SHRIMP facility became available at the Geological Survey of Canada and made it possible to obtain ages with high spatial resolution on a small number of single grains. At that point, McLelland et al. (2004) and Hamilton et al. (2004) initiated a series of SHRIMP U–Pb zircon investigations of the Adirondack AMCG suite. These results (Table 1) demonstrated that all major members of the Adirondack AMCG suite were emplaced within error of each other at 1155 ± 8 Ma and are basically coeval (Fig. 3). As can be seen in Table 1, SHRIMP ages are more consistent and reliable for these rocks than results of multigrain TIMS analyses, and we have used them wherever possible.

Uranium–Pb zircon ages for anorthosites now exist elsewhere in the Grenville Province (e.g., Higgins & van Breeemen 1992, Hébert & van Breeemen 2004, Morisset et al. 2009). These ages not only clarified issues regarding petrology; they also provided an essential framework enabling AMCG suites to be placed within the context of regional tectonic settings, as was subsequently done by Martignole (1996), McLelland
et al. (1996) and Corrigan & Hanmer (1997). It is the purpose of this paper to present this evidence together with the conclusion that rather than being "anorogenic", most AMCG magmatism took place within the context of regional orogenic settings. However, this does not rule out extensional, transtensional, or hotspot settings where these settings can be shown to be late- to post-tectonic. The key to such demonstration appears to lie with thorough, precise geochronology and field relationships. A detailed discussion begins with Adirondack massif anorthosite, with which the authors are most familiar.
The Adirondack Mountains are a southwestern extension of the Grenville Province to which they connect via the Frontenac Arch and Thousand Islands region of the St. Lawrence River (Fig. 2). The region is topographically divided into the Adirondack Highlands and Lowlands, separated by the Carthage–Colton Shear Zone (CCZ, Fig. 4). The former is underlain largely by orthogneiss metamorphosed to the granulite facies, and the latter by upper-amphibolite-grade metasedimentary rocks, notably marble. Both sectors have experienced multiple deformations resulting in refolded major isoclines. The Carthage–Colton Zone dips ~45° NW, and the latest displacement along it (ca. 1050 Ma) was down to the northwest, juxtaposing the Lowlands and Highlands (Buddington 1939, 1969, Johnson et al. 2004, Streepey et al. 2001, Selleck et al. 2005, Baird et al. 2008).

The major events of Adirondack tectonic evolution are summarized in Figure 5. The oldest rocks recognized in the region are calc-alkaline tonalites and granodiorites dated at ca. 1350–1250 Ma and exposed in the southern and eastern Adirondack Highlands (McLelland & Chiarenzelli 1990). Rocks of similar composition and age have been recognized in the Mount Holly complex (Fig. 4) of Vermont (Ratcliffe et al. 1996), the Proterozoic of western Connecticut (Walsh et al. 2004), and the New Jersey Highlands (Volkert & Aleinikoff 2005).

### TABLE 1 GEOCHRONOLOGY OF THE ADIRONDACK AMCG SUITE

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### Anorhotis and Olivine Gabro

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In addition, calc-alkaline plutonic rocks ranging in age from 1450 to 1300 Ma occur within the western Central Metasedimentary Belt of the Grenville Province (D, Fig. 2), which includes all of the Allochthonous Monocyclic Belt, except for the Adirondack Highlands and Morin terranes (Fig. 2). Dickin (2000) and Rivers & Corrigan (2000), among others, interpreted these calc-alkaline plutons as part of a southeast-facing Andean-style arc developed along the Laurentian margin during the interval ca. 1450–1300 Ma. Tectonic slices of this arc are preserved at the western margin of the Central Metasedimentary Belt Boundary Zone (D, Fig. 2) as well as in the composite Elzevirian arc (Fig. 5A), and are referred to as the Dysart suite (Easton 1992, Dickin & McNutt 2007). McEachern & van Breemen (1993) and Hanmer et al. (2000) proposed that the 1400–1300 Ma tonalitic rocks of the southern Adirondack Highlands and the Mount Holly complex of the Green Mountains, in Vermont (Fig. 5A, AD–GM), represent remnants of this arc rifted from Laurentia during formation of the Central Metasedimentary Belt back-arc basin, and referred to them as the Dysart – Mount Holly suite (Rivers & Corrigan 2000). From 1280 to 1220 Ma, these remnants served as nuclei for outboard arc magmatism (Fig 5A). At ca. 1220 Ma, the composite Elzevirian arc underwent accretion to Laurentia along the long-lived Central Metasedimentary Belt Boundary Zone, shutting off Elzevirian magmatism as the subduction zone stepped eastward (Fig. 5B). This sequence of events, as well as the configuration of terranes in Figure 5, differs from those of Carr et al. (2000) with respect to the placement of the Adirondack Lowlands, which is here considered to represent the trailing edge of the Elzevirian-age Central Metasedimentary Belt (Fig. 5A). This assignment is based on the presence of ca. 1210–1200 Ma calc-alkaline plutonic rocks of the Antwerp–Rossie suite (Figs. 2, 5B) within the Lowlands, together with oxygen isotope evidence that these plutons were emplaced above a west-dipping subduction zone (Peck et al. 2004), as shown in Figure 5B. Note that these rocks are absent from the Highlands, consistent with its separation from the Lowlands at that time. Recently, Chiarenzelli et al. (2010) have reported the discovery of an ophiolitic peridotite with oceanic crust affinities in the northeastern Lowlands (X, Fig. 2). These rocks have Nd model ages of ca. 1400–1350 Ma and contain metamorphic zircon dated at ca. 1214 Ma. This important result is wholly consistent with the tectonic model presented above.

The Frontenac terrane northwest of the St. Lawrence River is structurally separated from the Lowlands by the late Ottawan, steeply dipping Black Creek Shear Zone (Fig. 4), which the ca. 1160 Kingston dikes of Canada do not cross. Although similarities exist between the Frontenac and Lowlands terranes, there thus may be a significant discontinuity between them. Among the similarities between the two regions is the fact that both exhibit high-grade Shawinigan metamorphism (~650°C < T < 800°C, ~6.5 < P < 8 MPa), but no thermal Ottawan effects exceeding the closure temperature of titanite (~650°C); local 40Ar/39Ar hornblende ages indicate that Ottawan temperatures did not exceed ~500°C (Mezger et al. 1991, Streepey et al. 2001, Heumann et al. 2006). These results have been interpreted to indicate that during peak Shawinigan contraction (ca. 1190–1160 Ma), the Lowlands were thrust over the Highlands along a suture zone coincident with the present Carthage–Colton zone (Fig. 4E). The Diana quartz syenite complex intruded the suture zone at 1164 ± 11 Ma, and its oldest kinematic indicators are consistent with thrust displacement to the east (Baird 2008). At 1050 Ma, reactivation of normal faults along the Carthage–Colton zone resulted in topside down-to-the-west displacement (Fig. 4F), as recorded by inelastic
indicators in the ca. 1050 Ma Lyon Mountain granite, that locally intruded the zone (Selleck et al. 2005). The Frontenac and Lowlands terranes are interpreted as parts of the Orogenic Lid of Rivers (2008a).

**The Shawinigan orogeny**

At ca. 1200 Ma, the Adirondack Highlands – Green Mountain terrane was accreted to the Laurentian margin (Fig. 5C), and the early, contractional phase of the Shawinigan orogeny ensued. Deformation included the formation of early (F1) nappes, lithospheric thickening, high-grade metamorphism, and the emplacement of the synkinematic Hermon (ca. 1190 Ma) and Hyde School (ca. 1170 Ma) granitoids (Fig. 5C). Simultaneously, the Lowlands, at the leading edge of the upper plate, were thrust eastward over the Highlands along an early, west-dipping precursor Carthage–Colton Zone (Fig. 5C). During the thrusting, the precursor Carthage–Colton Zone was intruded by the 1164 ± 11 Ma Diana quartz syenite, which contains numerous early kinematic indicators showing top-side-up-to-the-east (Baird 2008, Baird et al. 2008). Following accretion, delamination of overthickened lithosphere took place and influxes of asthenosphere led to underplating of the crust by gabbroic magma and the evolution of late Shawinigan AMCG suites (Figs. 5C, D).

Corrigan (1995) appears to have been the first to recognize ca. 1190–1140 Ma Shawinigan tectonism during investigations near the small town of that name northeast of Quebec. At about the same time, Friedman & Martignole (1995), Martignole & Friedman
(1998) and Martignole (1996) reported widespread ca. 1190–1140 Ma metamorphism and tectonism within the Allochthonous Monocyclic Belt (Fig. 1). Subsequently, Rivers (1997) and Rivers & Corrigan (2000) defined the Shawinigan orogeny as encompassing the interval 1190–1140 Ma. However, both Martignole & Friedman (1998) and Martignole (1996, and references therein) stressed that contractional orogeny had largely terminated by 1160 Ma, and continuing metamorphism was due to thermal influxes from ca. 1160–1140 Ma AMCG.

plutons (e.g., the Morin suite at 1155 ± 3 Ma: Doig 1991). Wodicka et al. (2004) utilized U–Pb SHRIMP analysis analyses to date the post-tectonic monzonitic–gabbroic Chevreuil suite at ca. 1170–1160 Ma, thus post-dating regional ca. 1200–1180 Ma granulite-facies metamorphism in the Allochthonous Monocyclic Belt northwest of the Morin anorthosite (Fig. 1). Corrigan & Hanmer (1997) stressed that, “…evidence for deformation related to contractional tectonics appeared to be restricted to the interval 1.19–1.16 Ga”. In a related way, the undeformed ca. 1160 Ma Kingston dike swarm cross-cuts deformed 1176–1162 plutons and 1180–1160 Ma granulite-facies assemblages within the Frontenac terrane (Marcantonio et al. 1990). Within the Adirondack Lowlands, titanite and monazite ages of 1156 ± and 1171± Ma, respectively, exhibit no resetting, and no post-1172 Ma regional tectonism or metamorphism is observed (Mezger et al. 1991, Heumann et al. 2006). Local ca. 1155 Ma AMCG plutons exhibit minimal deformation outside of shear zones, and tectonism appears to have peaked at ca. 1180–1160 Ma. The large ca. 1170 Ma Maberly shear zone at the western margin of the Frontenac domain (Fig. 1) terminated thrust motion at 1156 ± 2 Ma (Corfu & Easton 1992, Rivers & Corrigan 2000). In summary, the Shawinigan regional contraction occurred throughout this entire area from ca. 1200 to 1160 Ma, whereas metamorphism from 1160 to 1140 Ma was largely limited to heat from AMCG suites or to local shear zones. Assuming that the Shawinigan orogeny in the Grenville Province continues to be assigned to the interval 1200–1140 Ma, it should always be emphasized that regional contraction ceased by ca. 1160 Ma. Within the southern Appalachians, Shawinigan contractional orogenesis does extend from ca. 1200 to 1140 Ma, and anorthosites are small, scarce, or absent (Tollo et al. 2006).

Following the Shawinigan orogeny, there does not appear to have been any significant regional deformation in the Adirondacks until ca. 1090 Ma when, after a brief interval (ca. 1103–1090 Ma) of granitic magmatism (the Hawkeye suite, McLelland et al. 1996), the entire Grenville Province began to experience effects of the Ottawa orogeny, considered to have been the result of collision of Laurentia and Amazonia (Figs. 5F, G), with the suture zone lying somewhere to the southeast of the present-day Appalachians. According to geothermometry, geobarometry, and seismic investigations, during the Ottawa orogeny, the Adirondack Highlands region attained a double thickness of crust and underwent granulite-facies metamorphism and F2 nappe emplacement (Bohlen et al. 1985, Valley et al. 1990, Kitchen & Valley 1995, McLelland et al. 1996). The major contractional phase of the Ottawa orogeny appears to have terminated by ca. 1050–1040 Ma, and minimally deformed plutons of Lyon Mountain granite (Figs. 4, 5G) were intruded at this time (McLelland et al. 2001). Locally, younger events of lesser magnitude continued to affect the region until ca. 950 Ma and are recorded by metamorphic zircon, titanite, and monazite. To understand how Adirondack anorthosites fit into this tectonic framework, we need to focus upon the results of recent geochronology for both the AMCG suite and the Shawinigan orogeny.

Details of the Shawinigan Orogeny in the Adirondacks

Evidence of Shawinigan plutonism and metamorphism is well developed in the Adirondack Lowlands, perhaps better so than elsewhere in the Grenville Province, and thus deserves some additional discussion. Figure 4 shows the presence of large quantities of igneous plutons in the Lowlands, and SHRIMP U–Pb zircon ages, as well as single-grain TIMS ages, demonstrate that all of these fall into the interval 1214–1155 Ma. The oldest rocks in the Lowlands consist of the Rossie diorite–quartz diorite suite and the Antwerp granitoid suite, consisting principally of biotite granite and granodiorite. Both units are strongly calc-alkaline. The Rossie diorite has been dated (L–ICP–MS) by Baird et al. (2008) at 1214 ± 38 Ma, and the Antwerp granitoid suite (SHRIMP), at 1210 ± 10 Ma by Wastenyes et al. (1999). The calc-alkaline biotite–hornblende Hermon megacrystic granite has been dated (SHRIMP) at 1182 ± 8 Ma (Heumann et al. 2006). Next in age is the ca. 1172 ± 2 Ma (single-grain TIMS) Ma calc-alkaline Hyde School gneiss, which consists of subequal amounts of pink leucogranite and gray tonalite (McLelland et al. 1992, Wastenyes et al. 1999). The granitic facies of the Hyde School gneiss is the same composition and age as the pink Rockport leucogranite that occurs along the St. Lawrence River, and the two are interpreted as belonging to a single, widespread Shawinigan event. Both the Hermon and Hyde School units have emplacement ages that coincide with metamorphic ages discussed in the following paragraph and are interpreted as syntectonic intrusions consistent with their fabrics (McLelland et al. 1992). Note that none of the ca. 1210–1170 Ma calc-alkaline rocks of the Lowlands are present in the Highlands, indicating that the two terranes were separate prior to this interval, and during much of it, but were accreted to one another by ca. 1155 Ma when AMCG stitching plutons were emplaced across both regions.

In a recent SHRIMP U–Pb zircon study of anatectic in migmatisitic metapelites of the Lowlands, Heumann et al. (2006) showed that partial melting took place at 1180–1160 Ma, and zircon grains exhibit no younger overgrowths. This result is consistent with the titanite data and demonstrates that the only high-grade metamorphism experienced by the Lowlands was at ca. 1210–1160 Ma (i.e., the Shawinigan orogeny). Shawinigan metamorphic ages occur in the Highlands as well (Heumann et al. 2006, Bickford et al. 2008), and xenoliths of high-grade Shawinigan garnet–sillimanite gneiss are preserved within an ~1106 Ma (J. Aleinikoff, pers.
commun. 2006, SHRIMP) metagabbro of the eastern Highlands (McLelland et al. 1988a). It is significant that Shawinigan-age regional Barrovian metamorphism in the Highlands was not documented until Heumann et al. (2006) and Bickford et al. (2008) obtained crystallization ages of 1180–1170 Ma for anatectic assemblages. These findings imply the following: 1) Shawinigan ca. 1200–1180 Ma plutons are common only within the Lowlands, and 2) there is extensive Ottawa granulite-facies metamorphism in the Highlands has overprinted earlier events. Similar considerations may have obscured recognition of Shawinigan metamorphism in parts of the eastern Grenville Province, and it would be instructive to obtain zircon ages and in situ monazite ages for leucosome in paragneisses older than ca. 1200 Ma. Importantly, Wodicka et al. (2003) have reported U–Pb zircon ages of 1170–1180 Ma for granitoid plutons in the vicinity of Havre St-Jean massif, and further dating is likely to reveal more of these ages.

When the 1156 ± 7 Ma average age (Table 1) of the AMCG suite is compared with the age of Shawinigan plutonism and anatexis in the Adirondacks (ca. 1210–1160 Ma), it is clear that the ages closely approach one another at 1160 Ma. In addition, the three oldest members of the granitoid suite have average ages exceeding 1170 Ma, and the next four oldest members are within error of mid-1170 Ma ages of the Shawinigan orogeny. Even the four oldest anorthosite samples are within error of ca. 1170 Ma. Given this age overlap, and the proximity of all of these rocks, it cannot be realistically supposed that the Adirondack AMCG suite is in any fashion “anorogenic”. Instead, the formation and evolution of the suite must be incorporated into a working model within which it can form and evolve in a setting that is late- to early post-tectonic. As we shall later argue, this type of environment applies to several major AMCG complexes.

A MODEL FOR THE GENESIS OF AMCG SUITES

The model is based upon the quantitative investigations of Houseman et al. (1981), Dewey (1988), Turner et al. (1992) and Platt & England (1994), among others, who have applied geophysical reasoning and calculations to orogeny, mountain building, and the late collapse of orogens. These authors concluded that contractional orogeny results in overthickened lithospheric roots, and that the latter are likely to be denser than the asthenosphere. A large portion of the overthickened lithosphere is then removed either by mechanical delamination or by convective removal of the thermal boundary-layer (Fig. 6). The delaminated lithosphere is replaced by asthenosphere that undergoes compression melting to produce gabbroic magmas that pond at the crust–mantle interface. During removal of the lithospheric root, the orogen rises in elevation as hotter, less dense material replaces the dense keel and produces buoyancy forces. The increased elevation results in increased potential energy of the plateau, which responds by exerting outwardly directed, extensional horizontal (F_h) that cause the underlying crust to push sideways until it balances whatever contractional forces (F_C) continue to exist and provide support for topographic elevations. At this point, the orogen may be said to have attained a state of mechanical equilibrium that supports mountainous topography and that simultaneously experiences collapse along low-angle normal faults until isostatic equilibrium is attained. It is estimated that the equilibrium state may last for 10–20 Ma (Platt & England 1994). During this interval, AFC processes may result in crust–mantle interactions recorded in geochemical signatures of trace elements (Emslie et al. 1994, Bickford et al. 2010).

The creation of a mechanically equilibrated state in the delaminated orogen is conducive to the development of an AMCG suite, because: 1) it remains tectonically stable for protracted periods of time, 2) it is associated with the ponding of large quantities of gabbro at the crust–mantle interface where hot, dry conditions exist, 3) it promotes the formation of coarse, commonly unzoned plagioclase of intermediate composition that floats in denser high-pressure melts, and 4) it facilitates the return of olivine and pyroxene cumulates back to the mantle. Finally, 5) the model provides the heat necessary to partially melt large volumes of deep continental crust. Although latent heat provides much of the required thermal energy, Platt & England (1994) showed that the convective removal of the lithospheric boundary-layer proceeds so rapidly that hot asthenosphere arrives at the base of the crust without having lost much heat by conduction, thus creating a thermal environment much like that associated with plumes or hot spots. In short, the late- to post-tectonic environment that exists in overthickened, delaminated orogens possesses all of the properties once thought to characterize “anorogenic” environments, and moreover, it is consistent with geochronological data and field evidence.

During the late- to post-tectonic phase of the Shawinigan orogeny, regional extension took place as contractional forces waned. Major manifestations of this extension are the Flinton and Twelve Mile basins (Fig. 5D), which host basinal sediments containing ca. 1155 Ma detrital zircon and which have been constrained to ca. 1150–1100 Ma in age (Sager-Kinsman & Parrish 1993, Wodicka et al. 1996).

AMCG MASSIFS BEYOND THE ADIRONDACKS

The Central Grenville Province

Figure 2 shows the major anorthosite massifs of the northeastern Grenville Province, as well as those situated north of the Grenville Front in Labrador. Of special importance to this discussion are the Morin (1153 ± 2 Ma, Doig 1991) and the Lac-St-Jean massifs
The CNadian Mineralogist (Higgins & van Breemen 1992, 1996, Higgins et al. 2002, Hébert & van Breemen 2004), which we discuss below. The Atikonak (1133 ± 10 Ma, Emslie & Hunt 1990) and the northern part of the Havre St. Pierre (1126 ± 7 Ma, Emslie & Hunt 1990) massifs remain problematic, because these ages are inferred from dating adjacent granitoids. However, Wodicka et al. (2003) have recently dated anorthosite in the northwest lobe of the Havre St. Pierre massif and obtained an age range of 1139–1150 Ma, within error of the Adirondack, Morin, and Lac-St-Jean AMCG suites.

As stated previously, Friedman & Martignole (1995), Martignole (1996), Corrigan & Hanmer (1997), Martignole & Friedman (1998) and Wodicka et al. (2004), among others, stressed that intrusion of the 1153 ± 3 Ma Morin AMCG suite postdates regional Shawinigan contractional orogeny (Fig. 7). Accordingly, it appears that the Morin anorthosite and related granitoids formed in a tectonic setting closely resembling that proposed for the Adirondacks (i.e., early post-tectonic delamination with influxes of fresh asthenosphere to the base of the crust, where the AMCG magmas evolved).

The entire Morin terrane was affected by Shawinigan contractional metamorphism from ca. 1190 to 1160 Ma, and the type locality of the Shawinigan orogeny lies at the southeastern edge of the Morin terrane ~250 km southwest of the enormous Lac-St-Jean massif. Although the Lac-St-Jean massif was displaced from its original site of intrusion, it seems unlikely that Shawinigan deformation and metamorphism did not affect the region into which the batholith was emplaced. Notwithstanding, there are currently no Shawinigan ages that have been reported for the area. However, as described earlier, ca.1190–1160 Ma Shawinigan metamorphism was not recognized in the Adirondack Highlands until SHRIMP analysis of zircon grains in the leucosome of migmatitic metapelites revealed its presence. A geochronological reconnaissance of similar lithologies in the central Grenville Province might well yield some Shawinigan ages and should be undertaken. Moreover, as described below, intrusion of the Lac-St-Jean batholith is demonstrably associated with large-scale, ductile strike–slip shear zones that manifest compressive forces at ca. 1157 Ma.

**Fig. 6.** Schematic representation of an overthickened collisional orogen undergoing lithospheric delamination and consequent orogen rebound and collapse along low-angle normal faults during late phases of orogenesis. Delamination represented by foundering of lithospheric keel; a convective erosion of the thermal boundary-layer is just as likely, however. AS refers to asthenosphere that ascends to the crust–mantle interface and undergoes decompression-induced melting to yield a gabbroic magma that fractionates olivine (ol) and pyroxene (px), which sink back into the mantle, and intermediate plagioclase (pgf), which floats. The plagioclase accumulates into a crystal mush (black squares) at the base of the crust. Ambient heat and heat of crystallization result in melting of the lower crust to yield mangeritic, charnockitic, and granitic magmas (MCG) of the AMCG suite. Orogenic contractional forces are represented by $F_C$, and the horizontal components of buoyancy forces, by $F_B$. 

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Dating of the large Lac-St-Jean anorthosite massif was undertaken by Hervet et al. (1994), Higgins & van Breemen (1992, 1996), Higgins et al. (2002), and by Hébert & van Breemen (2004). The results of these investigations (Table 2) are commonly cited as documenting a ~20 m.y. interval of crystallization from 1160–1140 Ma (e.g., Rivers 1997, Hébert & van Breemen 2004). However, examination of the data in Table 2 reveals that support for this interval is not compelling, and a much more restricted interval of crystallization is more likely. Specifically, there are six ages that directly date the crystallization of anorthosite or petrogenetically related rocks (samples 1, 2a, 3, 8, 10, and 13). Five of these six fall between 1157 and 1148 Ma with errors (2σ) of 2–4 Ma. The sixth (sample 13) is dated at 1140 ± 10 Ma. The average of all six samples is 1154 ± 4 Ma, which we consider to be the emplacement age of the anorthosite. It is an age that is coincident with the 1160–1140 Ma post-contractional phase of the Shawinigan event and with emplacement of the Marcy and Morin massifs (Fig. 7). This age is further supported by the observation that 1) the 1146 ± 3 Ma Lac LaBrèque granitic pluton is completely enclosed by the anorthosite, and 2) the Du Bras granophyric pluton cross-cuts the anorthosite. In addition, late anorthositic pegmatites within the Lac-St-Jean massif commonly contain a core of granophyre, two of which yield ages of 1155 ± 2 Ma and 1152 ± 2 Ma. Thus the anorthosite must be of this age or older. One granophytic core has been dated at 1142 ± 2 Ma, and was interpreted by Higgins & van Breemen (1992) as a late differentiate of the anorthositic magma, although no persuasive evidence is presented for this. Moreover, residual granophytic liquids are not likely to form from anorthositic parent magmas, which follow a strong iron-enrichment Fenner trend, leading to ferrodioritic residua such as in samples 1, 8, and 10. We suggest that the granophyres are crustal A-type partial melts produced by heat from underplated gabbro and the massifs themselves. Finally, we note that metamorphic zircon in the granophyres yields ages of 1142 ± 2 Ma, clearly demonstrating that this age does not reflect emplacement of the anorthosite. In fact, Higgins & van Breemen (1992) argued persuasively that field and textural evidence in the NNE-trending megadykes (samples 1, 2) demonstrate that they were emplaced while the anorthosite was still, in part, not fully crystallized. By the time that the younger of the two megadykes (Sample 2) had been emplaced, the anorthosite was undergoing solid-state strain. They used this evidence to argue for rapid emplacement (i.e., a one to two million year interval, not ~20 m.y.) of the anorthositic magma, with which we agree. We also agree that the steep NNE strike-slip faults (with releasing bends) that cut the anorthosite might well have served as important conduits for the Lac-St-Jean magmas to reach the middle crust from its staging area at the crust–mantle interface. Such faulting would be consistent with the late transition of the regional stress-

![Fig. 7. Chart showing the correlation between emplacement of major AMCG massifs of northeastern North America and the waning phase of orogeny. Black represents contractional orogeny, light gray represents magmatism, and dark gray represents pre-contractional arc magmatism.](image-url)
field from compression to regional extension. The ages of associated granitoids (samples 1, 9a, b, 11, 15) are consistent with the foregoing arguments.

South of the Abbe–Huard lineament in the Havre St. Pierre massif (Fig. 1), the Lac Allard and Magpie River lobes of the Havre St. Pierre anorthosite have been directly dated at ca. 1060 Ma (van Breemen & Higgins 1993, Morisset et al. 2009), which coincides with the terminal stages of the Ottawa orogeny, as emphasized by Morisset et al. (2009). In addition to the Havre St. Pierre example, Owens et al. (1994), Dymek & Owens (1998) and Owens & Dymek (2001, 2004) have mapped, analyzed, and dated a number of small bodies of ca. 1080–1010 Ma anorthosite and related ca. 1050 Ma A-type, post-orogenic granitoid plutons in the Parc des Laurentides area (Fig. 2) that lie in a NNE-trending belt (CRUML belt, Fig. 1) extending from Quebec City to the northeastern margin of the Lac-St-Jean massif. The ca. 1080–1010 Ma anorthosites of the CRUML Belt and the central Appalachians differ from the large massifs of the Adirondacks and Grenville Province in several important ways, e.g., they contain Fe3+-rich ilmenite with exsolution lamellae of hematite rather than ilmenite–magnetite, their antiperthitic plagioclase compositions are richer in sodium and potassium (An48–62Or2–3) than in the ca. 1150 Ma massifs (An23–40Or4–25), and they contain orthopyroxene, but neither clinopyroxene nor olivine.

### TABLE 2. GEOCHRONOLOGY OF THE LAC-ST-JEAN AMCG SUITE*

<table>
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<td>1) Lac Chabot ferrodiorite megacycle</td>
<td>1157 ± 3</td>
<td>Zm</td>
<td>Crystallization</td>
<td>Intruded into active sinistral NNE vertical shear zone</td>
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<tr>
<td>2a) Bégin leucocrotolite megacycle</td>
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<td>Bdl</td>
<td>Crystallization</td>
<td>Same as above. Deformation outlasts magnetism</td>
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</tr>
<tr>
<td>3) Anorthositic pegmatite</td>
<td>1156 ± 2</td>
<td>Bdl</td>
<td>Crystallization</td>
<td>Two concordant points</td>
</tr>
<tr>
<td>4) Granophyre core in anorthositic pegmatite</td>
<td>1142 ± 2</td>
<td>Zm</td>
<td>Crystallization</td>
<td>Age interpreted as the age of anorthosite crystallization</td>
</tr>
<tr>
<td>5a) Granophyre segregation in anorthosite</td>
<td>1155 ± 2</td>
<td>Zm</td>
<td>Crystallization</td>
<td>Age is NOT extrapolated to that of anorthosite crystallization</td>
</tr>
<tr>
<td>5b) Granophyre core in anorthositic pegmatite</td>
<td>1142 ± 2</td>
<td>Zm</td>
<td>Crystallization</td>
<td>200 m away from sample 7</td>
</tr>
<tr>
<td>6a) Small segregation of granophyre</td>
<td>1152 ± 2</td>
<td>Zm</td>
<td>Crystallization</td>
<td>Thermal recrystallization or lead loss</td>
</tr>
<tr>
<td>6b) Single large zircon crystal in granophyre</td>
<td>1142 ± 5</td>
<td>Zm</td>
<td>Metamorphic</td>
<td></td>
</tr>
<tr>
<td>7) Lac Labrèque granite pluton (5 × 12 km)</td>
<td>1146 ± 3</td>
<td>Zm</td>
<td>Crystallization</td>
<td>Completely enclosed in anorthosite, minimum age</td>
</tr>
<tr>
<td>8) Lac Kenogami ferrodiorite in contact zone</td>
<td>1143 ± 4</td>
<td>Zm</td>
<td>Crystallization</td>
<td></td>
</tr>
<tr>
<td>8a) Lac Kenogami farsandite</td>
<td>1150 ± 3</td>
<td>Zm</td>
<td>Crystallization</td>
<td></td>
</tr>
<tr>
<td>8b) A second fraction</td>
<td>1190</td>
<td>Zm</td>
<td>Crystallization</td>
<td></td>
</tr>
<tr>
<td>9) Coarse anorthosite</td>
<td>1140 ± 10</td>
<td>Zm</td>
<td>Crystallization</td>
<td>Three very discordant (12%) points</td>
</tr>
<tr>
<td>10) Coarse anorthosite</td>
<td>1163 ± 18</td>
<td>Tm</td>
<td>Metamorphism</td>
<td>At contact with Du Bras pluton</td>
</tr>
<tr>
<td>11) Lac Kenogami champokite – farsandite</td>
<td>1155–1135</td>
<td>Zm</td>
<td>Crystallization</td>
<td>Same as sample 9a</td>
</tr>
<tr>
<td>12) Du Bras granophyre pluton</td>
<td>1148 ± 2</td>
<td>Zm</td>
<td>Crystallization</td>
<td>Coeval with, similar to Lac Labrèque pluton</td>
</tr>
<tr>
<td>13) Coarse anorthosite</td>
<td>1140 ± 10</td>
<td>Zm</td>
<td>Crystallization</td>
<td></td>
</tr>
<tr>
<td>14) Wollastonite skarn</td>
<td>1148 ± 2</td>
<td>Zm</td>
<td>Crystallization</td>
<td>At contact with Du Bras pluton</td>
</tr>
<tr>
<td>15) Pyroxene monzonite</td>
<td>1148 ± 4</td>
<td>Zm</td>
<td>Crystallization</td>
<td>At contact with anorthosite</td>
</tr>
</tbody>
</table>

They are considered to have evolved under higher \( f(\text{O}_2) \) conditions and within somewhat more hydrous environments, and assimilated more continental crust than the large massifs of the Grenville Province. They have been referred to as alkalic anorthosites (e.g., Owens et al. 1994, Owens & Dyck 2001, 2004). In contrast to the alkalic anorthosites, the pre-1100 Ma anorthosite massifs are referred to as pyroxene–labradorite or pyroxene – olivine – labradorite anorthosites on the basis of their dominant mineralogy.

All of the ages cited in the previous paragraph fall into the late- to post-tectonic phase of the Ottawa orogeny (Fig. 7) and are approximately synchronous with late extension and collapse of the orogen at ca. 1060–1020 Ma (Rivers 2008a) making them good examples of the late- to post-tectonic setting for AMCG magmatism (Morisset et al. 2009). More specifically, peak and late phases of the Ottawa orogeny within the Adirondacks were followed by delamination, pop-up, and collapse, (accompanied (and enhanced) by the ascent of large volumes of ca. 1050 Ma Lyon Mt. leucogranite (Figs. 4, 5G), considered to be correlative with ca. 1060–1050 Ma Parc des Laurentides (Fig. 2) granitoids. The CRML Belt projects farther to the east of the Adirondacks, where it may lie beneath Phanerozoic cover. This belt re-emerges within the Proterozoic inliers of the Pennsylvania and Virginia Appalachians (i.e., the Montpelier anorthosite, 1045 ± 10 Ma, and the Roseland anorthosite, 1045 ± 44 Ma anorthosites: Aleinikoff et al. 1996, Pettingill et al. 1984). These small bodies are associated with ca. 1050 Ma granites and leucogranites similar to those in the CRUML Belt MCG granitoids and Adirondack Lyon Mountain Granite. Based upon these observations, we re-emphasize that Ottawa orogenesis was followed by late- to post-orogenic AMCG magmatism with the same basic timing and mechanisms as proposed for the late- to post-Shawinigan AMCG complexes. Differences between the alkali-rich anorthosites and the laboratory massifs can be accounted for by differences in fluids and magmatic interaction with crustal materials.

The AMCG Complexes complexes of eastern Labrador

Important massifs of olivine–pyroxene labradorite-type anorthosite, leucanorite, and leucotroctolite, together with MCG granitoids, occur in Labrador immediately north of the eastern Grenville Province (i.e., Michikamau, ca. 1460 Ma, Harp Lake, ca. 1448 – 1426 Ma, Mistastin, ca. 1420 Ma, mainly MCG, and Nain, 1350–1290 Ma complexes). These undeformed and pristine Early to Middle Elsonian (1.46–1.29 Ga) massifs have been regarded as type examples of “anorogenic” AMCG magmatism and exhibit characteristic large crystals of intermediate plagioclase and aluminous pyroxene (Emslie 1978, 1985, Emslie et al. 1994). The Nain complex is localized within a steep Paleoproterozoic shear zone (the Torngat orogen) between the Churchill and Nain provinces, and the other three lie just to the southwest. Recent geochronological, petrological, and tectonic reviews by Gower (1996), Gower & Krogh (2002) and Myers et al. (2008) have opened a new chapter in our understanding of these AMCG complexes and are summarized in this section.

Utilizing both detailed field work and zircon dating, Gower & Krogh (2002) have demonstrated that the eastern Grenville Province experienced strong deformation, metamorphism, and plutonism during the ca. 1.52–1.46 Ma Pinwarian orogeny, whose Andean-type plate margin bordered southeastern Laurentia directly south of the future sites of the early Elsonian AMCG suites in Labrador. Terminal Pinwarian events (ca. 1.46 Ga) coincide in time with the onset of AMCG magmatism (ca. 1460 Ma) in the Harp Lake, Michikamau, and Mistastin massifs (Fig. 7). In addition, the Shabogamo and Michael gabbros, located south of the massifs and mostly within the Grenville Province (Fig. 2), are dated at 1459 ± 23 Ma and 1472 ± 27 Ga, respectively, and manifest extension overlapping and following Pinwarian contraction. Gower & Krogh (2002) accounted for the Pinwarian orogeny by proposing a north-dipping subduction zone that was active from 1520 to 1460 Ma. Subsequently, orogeny and magmatism shut off along the continental margin, and the extension-related 1469–1426 Ma Shabogamo and Michael gabbros (Fig. 2) were emplaced in proximity to the present-day Grenville Front. At about the same time (1460–1420 Ma), the Harp Lake, Michikamau, and Mistastin AMCG complexes were emplaced in pulses that grew younger from south to north.

In order to account for these observations, Gower & Krogh (2002), following Gutscher et al. (2000), advocated flat-slab subduction involving an overridden spreading ridge that provided thermal buoyancy for the lower plate. This proposal explains the shutting off of Pinwarian magmatism at the subduction zone as well as the south-to-north younging of intrusive ages. Gower & Krogh (2002) suggested that the spreading ridge not only provided buoyancy, but might also have fed asthenospheric gabbro through the overlying lithosphere to the base of the continental crust, where it ponded and fractionated. Gower & Krogh (2002) also noted that the “entire region underpinned by flat-subducted crust is one of tectonic instability and pockets of magmatism might be anticipated in any weak spot”. We expand this suggestion to include delamination or breakoff of the cooling lithosphere(s), resulting in the ascent of deeper, hotter asthenosphere and decompression melting, producing gabbro that ponded at the base of the crust. Another variation is that the narrow wedge of asthenosphere above the leading tip of the now-dehydrated oceanic crust of the lower plate was replenished by convective circulation providing fresh, hotter asthenosphere that underwent decompression melting to yield the underplated gabbros. Although difficult to
demonstrate at present, the flat-slab model is attractive and consistent with what is observed at the surface for the Harp Lake, Mistastin, and Michikamau plutons as well as the Michael–Shabogamo dike swarms (Fig. 1). Moreover, it is consistent with a late- to post-tectonic setting (Fig. 7). Rivers & Corrigan (2000) and Rivers (2008b) have proposed that the AMCG magmatism of Labrador is the consequence of back-arc extension, which may be compatible with the flat-slab model described here.

It is difficult to reconcile the flat-slab model with the Nain complex, which is 74 million years younger than the neighboring AMCG plutons. However, the recent investigations of Myers et al. (2008) has shed considerable light on the origin of the Nain complex, which we address in the next to last section of the paper.

Anorthosites and AMCG suites of southern Norway

The 1100–930 Ma (Bingen et al. 1993) Sveconorwegian Orogen bears resemblances to the Grenville Province with which it was presumably contiguous during much of mid-Proterozoic time; it also was affected by deformation analogous to the ca.1100–1030 Ma Ottawan (i.e., Grenvillian) orogeny. During the past two decades, a great deal of high-quality geochronological data have emerged on the Sveconorwegian region and has served to clarify important details of its evolution. We note that the western portion of the orogen is allochthonous and has been transported and overprinted by metamorphism in the Jotun Nappe Complex of southwestern Norway. The eastern part of the orogen is located on the southwestern tip of Norway and reappears in eastern Sweden. Anorthosites and AMCG suites exist in both sectors. In the Jotun Nappe portion of the orogen, regional convergence and metamorphism appear to have begun at ca. 1110 Ma and were followed by subduction beneath Btica from ca. 1050 to 1040 Ma (Bingen & van Breemen 1998). Subsequently, thrusting migrated eastward. Relaxation occurred at about 970 Ma, and a massif anorthosite was emplaced at that time (Lundmark & Corfu 2008, Bingen et al. 2005). Farther southeast, regional metamorphism extended until 930 Ma when the very large Rogaland anorthosite complex, including the Bjerkreim–Sknadal layered leucocratic lopolith, was rapidly emplaced (Schärer et al. 1996, Bingen et al. 2005). All of these investigators remarked on the barely post-tectonic emplacement of these AMCG massifs and their relationship to an extensional, collapse phase of the Sveconorwegian Orogeny.

Miscellaneous examples of AMCG complexes in lithospheric-scale shear zones

Myers et al. (2008) have recently published the results of a thorough investigation of the Nain igneous plutonic complex (Figs. 2, 7) that must be part of any discussion of the tectonic setting of AMCG suites. Here we summarize the results of their work. The ~300 × 75–100 km Nain batholith was intruded (1363 – 1289 Ma) into the ~100 km-wide Torngat orogen (Hamilton 2008) that marks the suture between the Churchill and Nain provinces (Fig. 2). The Nain complex consists of large coalescing and cross-cutting plutons of anorthosite, monzonitic, ferrodioritic, troctolitic, and granitic rocks that intrude into sharp, well-defined fractures and give little support for diapiric models of intrusion. Ring complexes, cauldron subsidence, dikes and sheets characterize the intrusive modes. Emplacement occurred in several distinct pulses of AMCG magmatism spread over 75 million years. The origin of the Nain complex is closely related to the ~1000 × 50 km lithospheric-scale Gardar–Voissey Bay Fault Zone that trends east–west from the Gardar Province, Greenland (Upton et al. 2003) and intersects the NW–SE Torngat orogen at a high angle. The fault zone is steep, exhibits sinistral offset of ~20 km, and contains AMCG and other intrusions that have been dated and constrain the timing of displacement. The ages of magmatism and displacement correlate closely with the five 1363–1289 Ma intrusive pulses that built the Nain complex from 1363 to 1289 Ma. The interpretation is that transtensional displacement on the fault zone reactivated the Torngat suture causing fracturing, pressure drops, and potential openings that led to decompression melting in the mantle and provided conduits for gabbroic magma.

Emplacement of AMCG suites into ancient, lithospheric-scale shear belts and suture zones has been demonstrated in several areas other than the Nain Complex, e.g., the Laramie and Horse Creek complexes (Fig. 1) in Wyoming (Scoates 1994, Scoates & Chamberlain 1995, 1997), as well as the Phanerozoic (410–420 Ma) AMCG complexes in Air, Niger (Moreau et al. 1994). The Wyoming examples occur within the Cheyenne Belt that marks the suture between the Archean Wyoming Province and the Proterozoic Colorado Province. In the case of the Horse Creek anorthosite, Scoates & Chamberlain (1995, 1997) have dated its emplacement at ca. 1.76 Ga, which coincides with the late stages of the Medicine Bow orogeny, which involved a local transtensional zone along the suture that served as a conduit for AMCG magmas (Frost et al. 2000). A similar origin may apply to the ca. 1.43 Ga Laramie anorthosite complex (Frost et al. 1990, 2000, Scoates & Chamberlain 1995, 1997), but the tectonic setting at that time remains uncertain.

Although not Proterozoic, the ca. 410–420 Ma subvolcanic AMCG ring-dike complexes of the Air region in Niger (Brown et al. 1989, Demaiffe et al. 1991, Moreau et al. 1994) deserve mention. The belt was once thought to form a N–S hot-spot track, ~1200 km long. However, more recent dating indicates that consistent younging to the south does not exist, and most workers have abandoned the hotspot model.
Instead, it now appears likely that the belt formed at ca. 410–420 Ma in response to reactivation along the ~1000 km long, N–S trending, dextral strike–slip Raghane lithospheric-scale shear zone that gave rise to a complex array of N20–50°E Reidel shears that served as magma conduits (Moreau et al. 1994).

**SUMMARY AND DISCUSSION**

Detailed geochronology in the Grenville Province, Labrador, and southern Norway indicates that significant examples of AMCG magmatism are closely correlated with the late- to post-tectonic stages of orogeny or with strike–slip displacement along lithospheric-scale fault zones and reactivated suturets, or both (Fig. 7). There does not appear to be support for a truly “anorogenic” emplacement. The most significant factor in the genesis of Proterozoic massif anorthosite is delivery of asthenosphere to the base of thick, stable continental crust where decompression melting yields a gabbroic magma. These melts undergo slow, high-pressure, anhydrous fractionation and are replenished by fresh influxes of gabbroic magma. Related melting of the lower crust produces felsic members of the AMCG suite, and batches of these granitoids eventually ascend, along with early, high-pressure anorthositic plagioclase cumulates contained within magmas of broadly leucogabbroic to leuconoritic composition. Following emplacement in the middle to upper crust, lower-pressure fractionation takes place in the proto-anorthositic magmas. Detailed U–Pb zircon geochronology has established that this polybaric evolution occurs in late- to post-tectonic settings. Although precise mechanisms are difficult to prove, several candidates seem plausible and are consistent with evidence. These include delamination of overthickened lithosphere in contractional orogens, flat-slab subduction, perhaps accompanied by delamination, extension in back-arc basins, and slab breakoff. What all of these have in common is the ability to deliver asthenosphere to the base of the crust and to do so within the context of late- to post-tectonic conditions. A special subclass of these settings is that of the transtensional reactivation of lithospheric-scale shear zones, especially those involving the reactivation of ancient sutures. In the examples investigated to date, there are no good examples of AMCG suites generated by hotspots or plumes.

In retrospect, we propose that the localization of AMCG magmatism in North America along a relatively narrow belt is due to the fact that the belt represents a protracted (ca. 1800–1000 Ma) orogenic continental margin arc system ~4000 km long that was active along the southeastern margin of Laurentia (Anderson 1983, Rivers & Corrigan 2000, Karlstrom et al. 2001). Within this context, late- to post-tectonic magmatism could produce AMCG complexes *via* the mechanisms presented in this paper. However, by itself, this proposal does not explain why equivalent AMCG complexes of Neoproterozoic age have not been discovered. On the other hand, the Phanerozoic AMCG complexes of Air document that these rocks were not restricted to the Proterozoic. The Air examples are subvolcanic intrusions, whereas AMCG suites developed within compressional belts are emplaced at mid-crustal depths and require longer intervals of erosion to arrive at the surface.

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