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METAMORPHISM OF THE CANADIAN SHIELD, ONTARIO, CANADA. II. PROTEROZOIC METAMORPHIC HISTORY

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Abstract

This paper is a complement to the Tectonometamorphic Map of the Canadian Shield, and contains a brief description of the metamorphic history of Proterozoic orogenic belts and rock sequences in the Canadian Shield in Ontario, with a focus on rocks of the Southern and Grenville provinces. Metamorphism in the Southern Province has generally been attributed to the Penokean orogeny (~1879–1820 Ma), although the timing of this metamorphism is poorly constrained by either relative or absolute ages. In the Sault Ste. Marie – Sudbury area, previously described regional Na- and K-metasomatism has likely altered the original assemblages of metamorphic minerals, making unravelling of the metamorphic history of these rocks problematic. Metamorphism appears to postdate the emplacement of the Sudbury Igneous Complex at 1850 Ma; however, it is unclear if regional metamorphism in the Sault Ste. Marie – Sudbury area is related to the early (1870–1820 Ma) or late-stage (1740–1700 Ma) events of the Penokean orogeny. Three main metamorphic events are associated with the Grenville (1300–950 Ma) orogeny. These events are not necessarily orogen-wide in their effects. The first, at 1250–1240 Ma, affected rocks of the Composite Arc Belt, and is similar in style to that found in Superior Province granite–greenstone belts. It consists of a dominant subgreenschist- to greenschist-facies event, with amphibolite-facies assemblages developed near larger plutonic bodies. The next event (1190–1170 Ma) is restricted mainly to the southern part of the orogen (southern Laurentian margin, Composite Arc and Frontenac–Adirondack belts), and is responsible for regional upper-amphibolite- to granulite-facies metamorphism. In parts of the orogen, metamorphic rocks formed at this time are well preserved (*e.g.,* Parry Sound and Frontenac domains), whereas in other areas (*e.g.,* Adirondack Highlands), the rocks are overprinted by younger, high-grade metamorphism. Pan-Grenville greenschist- to granulite-facies metamorphism in Ontario in the period 1070–1050 Ma was associated with major thrusting. In deeper structural levels, this event persists to 990 Ma. Tectonic unroofing after 1070 Ma likely played a major role in the current distribution of metamorphic rocks within the Grenville Province in Ontario.

Keywords: Ontario, Proterozoic, Grenville Province, Southern Province, metamorphism, granulite, amphibolite, greenschist, subgreenschist, Keweenawan Supergroup.

Sommaire

Cet article se veut un complément de la nouvelle carte tectonométamorphique du Bouclier Canadien; il contient un bref exposé de l'évolution métamorphique des ceintures orogéniques d'âge protérozoïque et des séquences du Bouclier Canadien en Ontario, avec une emphase sur les provinces géologiques du Sud et du Grenville. Dans la Province du Sud, on attribue généralement le métamorphisme à l'orogénèse pénokéenne (~1879–1820 Ma), quoique l'âge exact de cet épisode, fut-il mesuré par méthodes relatives ou absolues, demeure méconnu. Dans la région de Sault Ste. Marie – Sudbury, un épisode de métasomatose régionale sodique et potassique semble avoir modifié les assemblages originels de minéraux métamorphiques, rendant ainsi obscur le bilan des événements métamorphiques qui ont affecté ces roches. Le métamorphisme semble postérieur à la mise en place du complexe igné de Sudbury, à 1850 Ma. Toutefois, il y a ambiguïté à savoir si la recristallisation métamorphique régionale dans la région de Sault Ste. Marie – Sudbury date du stade précoce (1870–1820 Ma) ou tardif (1740–1700 Ma) de l'orogénèse pénokéenne. Par contre, trois événements principaux sont associés à l'orogénèse grenvillienne (1300–950 Ma). Ils n'ont pas nécessairement eu une influence à l'échelle de la ceinture orogénique entière. Le premier, à 1250–1240 Ma, a affecté les roches de la ceinture de l'Arc Composé, et ressemble à l'événement qui a affecté les ceintures de granite – roches vertes de la Province du Supérieur. Il a causé une recristallisation dominante aux conditions du faciès schistes verts, voire même sous-schistes verts, les assemblages typiques du faciès amphibolite étant développés près des massifs plutoniques importants. Le deuxième événement (1190–1170 Ma) est limité surtout à la partie sud de l'orogénèse (bordure sud du socle laurentien, Arc Composé et ceintures de Frontenac et de l'Adirondack), et serait responsable d'un métamorphisme régional aux conditions du faciès amphibolite supérieur ou du faciès granulite. Dans certaines parties de l'orogénèse, les roches affectées par ce stade sont bien conservées (par exemple, dans les domaines de Parry Sound et de Frontenac), tandis qu'ailleurs (par exemple, le plateau de l'Adirondack), les roches sont reprises par un épisode de métamorphisme intense plus jeune. Un métamorphisme pan-grenvillien aux conditions allant de faciès

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schistes verts à granulite en Ontario (1070–1050 Ma) a accompagné une période de chevauchement important. Aux niveaux inférieurs, cet événement a persisté jusqu'à 990 Ma. Un soulèvement tectonique postérieur à 1070 Ma aurait joué un rôle déterminant dans la distribution actuelle des roches métamorphiques de la Province du Grenville en Ontario.

(Traduit par la Rédaction)

Mots-clés: Ontario, protérozoïque, Province du Grenville, Province du Sud, métamorphisme, granulite, amphibolite, schistes verts, sous-schistes verts, Supergroupe de Keweenaw.

Introduction

This paper is an outgrowth of the compilation of metamorphic information for the Canadian Shield in Ontario, part of Ontario's contribution to the Tectonometamorphic Map of the Canadian Shield (Berman, in prep.). In this paper, I discuss the history of Proterozoic orogenic belts and rock sequences in the Canadian Shield in Ontario; this contribution is a companion to Easton (2000), in which I review the Archean metamorphic history.

This paper is divided into two parts. In the first part, I briefly describe the metamorphic history of Proterozoic rocks and orogenic belts in the Canadian Shield in Ontario. This section is not designed to present a comprehensive history of metamorphic studies in a particular area, but rather, to explain how units are portrayed on the map and to point out limitations in the data. The second part is a discussion of the constraints that metamorphic history places on the tectonic evolution of the Southern and Grenville provinces in Ontario. In brief, this paper and Easton (2000) can be considered as a set of "marginal notes" for the Ontario portion of the Tectonometamorphic Map of the Canadian Shield (Berman, in prep.).

Figure 1 illustrates the tectonic divisions of Ontario used in this study. Figure 2 is a scaled-down version of the metamorphic map covering the Southern and Grenville provinces. Details of the compilation process are given in Easton (1999b), with details of the legend design and conventions about terminology presented in Berman *et al.* (2000). In brief, the legend is akin to that used on the 1978 Metamorphic Map (Fraser *et al.* 1978) in that it portrays units based on metamorphic facies (subgreenschist, greenschist, amphibolite, granulite) and superposition of metamorphic events. Although several types of metamorphism can be identified in the better-studied parts of the Canadian Shield (*e.g.*, seafloor metamorphism, contact metamorphism, regional metamorphism, burial metamorphism, *etc.*), the Tectonometamorphic Map of the Canadian Shield emphasizes the distribution of regional metamorphic events.

A digital version of the 1:2 000 000-scale geological map of the Grenville Province (Davidson 1998b) was used as the geological base-map for the metamorphic map compilation covering Ontario, New York State and western Quebec. This base was selected because it provided seamless coverage across political boundaries, and because the Ontario portion does not differ significantly from the Tectonic Map of Ontario. During compilation, the base map was not modified, although new results of mapping projects were taken into consideration.

Definition of low-grade regional metamorphism

Defining the transition from diagenesis to regional metamorphism, *i.e.*, separating unmetamorphosed from metamorphosed rock, is not always clear, especially in clastic rocks in low-pressure environments, as such rocks may not develop diagnostic minerals such as zeolites (Frey 1987). This distinction can be further complicated by the effects of post-burial fluid flow, intense post-depositional chemical weathering, and regional hydrothermal alteration. The approach used herein follows that of Frey & Kisch (1987) and Turner (1981), in that processes are considered *metamorphic* where the bulk of the rock, including coarse particles of sand, is substantially affected, and where the presence of incipient foliation indicates the simultaneous yielding of the rock to stress-induced flow. Index minerals of the subgreenschist facies are zeolite, prehnite, pumpellyite and pyrophyllite (Turner 1981, Frey 1987). *Diagenesis* is used in the sense of "early diagenesis", in that it only includes changes taking place in a sediment between sedimentation and the completion of lithification or cementation.

In Ontario, many of the rock units that fall into the *subgreenschist facies* category were studied prior to the advent of modern methods for studying such rocks [*e.g.*, determination of "crystallinity" indices and polytypes, textural classifications, *etc.*; see Frey (1987) for details], thereby rendering the distinction between diagenesis, alteration, and metamorphism difficult. For the purpose of the map legend, no distinction is made between areas subjected to regional low-grade alteration and those subjected to low-grade regional metamorphism. These distinctions are discussed below under the relevant descriptive sections of this paper.

A further difficulty with respect to subgreenschist-facies units in Ontario was determining the timing of regional metamorphism. Where an areally extensive unit is subjected to regional prograde metamorphism, such as the Huronian Supergroup, the timing of metamorphism in the higher-grade zones can be used as a reasonable approximation of the timing of lower-grade

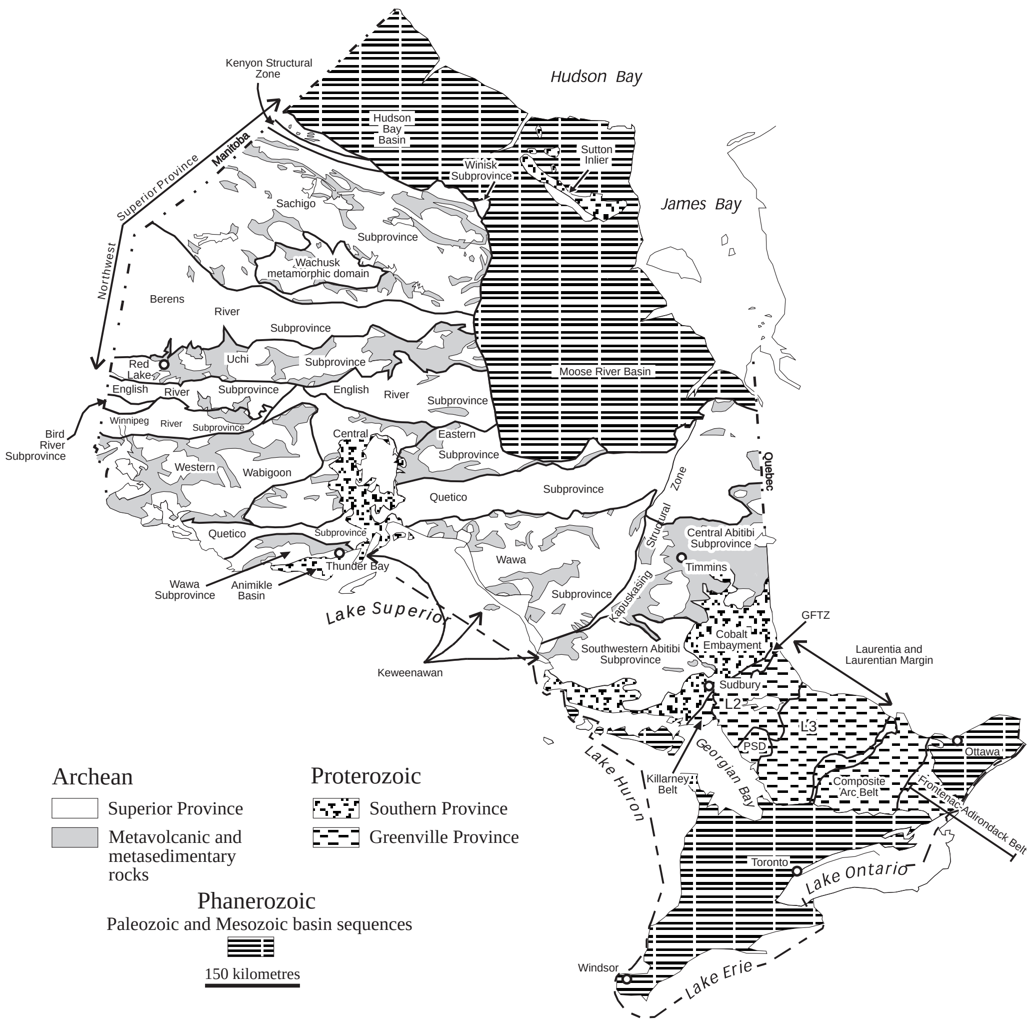


FIG. 1. Tectonic subdivision of Ontario used in this paper (*after* Thurston 1991).

metamorphism. K–Ar and Rb–Sr ages may also provide some insight into the timing of regional metamorphism, although in many cases in Ontario, the age determinations available are generally of poor resolution, or were collected for other purposes.

Metamorphic History of Proterozoic Rocks in Northwestern and Central Ontario

Sutton Inlier

Bostock (1971) reported the presence of chlorite, white mica and minnesotaite in sedimentary rocks of the Paleoproterozoic Sutton Group, located in the Sutton Inlier west of James Bay (Fig. 1, 2). Gabbro bodies in the inlier contain epidote and chlorite, and Na-metasomatized feldspar containing fine-grained white mica (Bostock 1971). The above assemblages indicate subgreenschist-facies metamorphism. Penetrative recrystallization of the gabbro bodies is interpreted to indicate a regional metamorphic event. The gabbros locally heated their sedimentary country-rocks, forming tremolite or stilpnomelane. The age of metamorphism is unknown, but is most likely related to the Trans-Hudson orogen to the north. Following Bostock (1971) and Fraser *et al.* (1978), metamorphism of the Paleoproterozoic rocks of the Sutton Inlier is assigned to Geon 18 (Fig. 2).

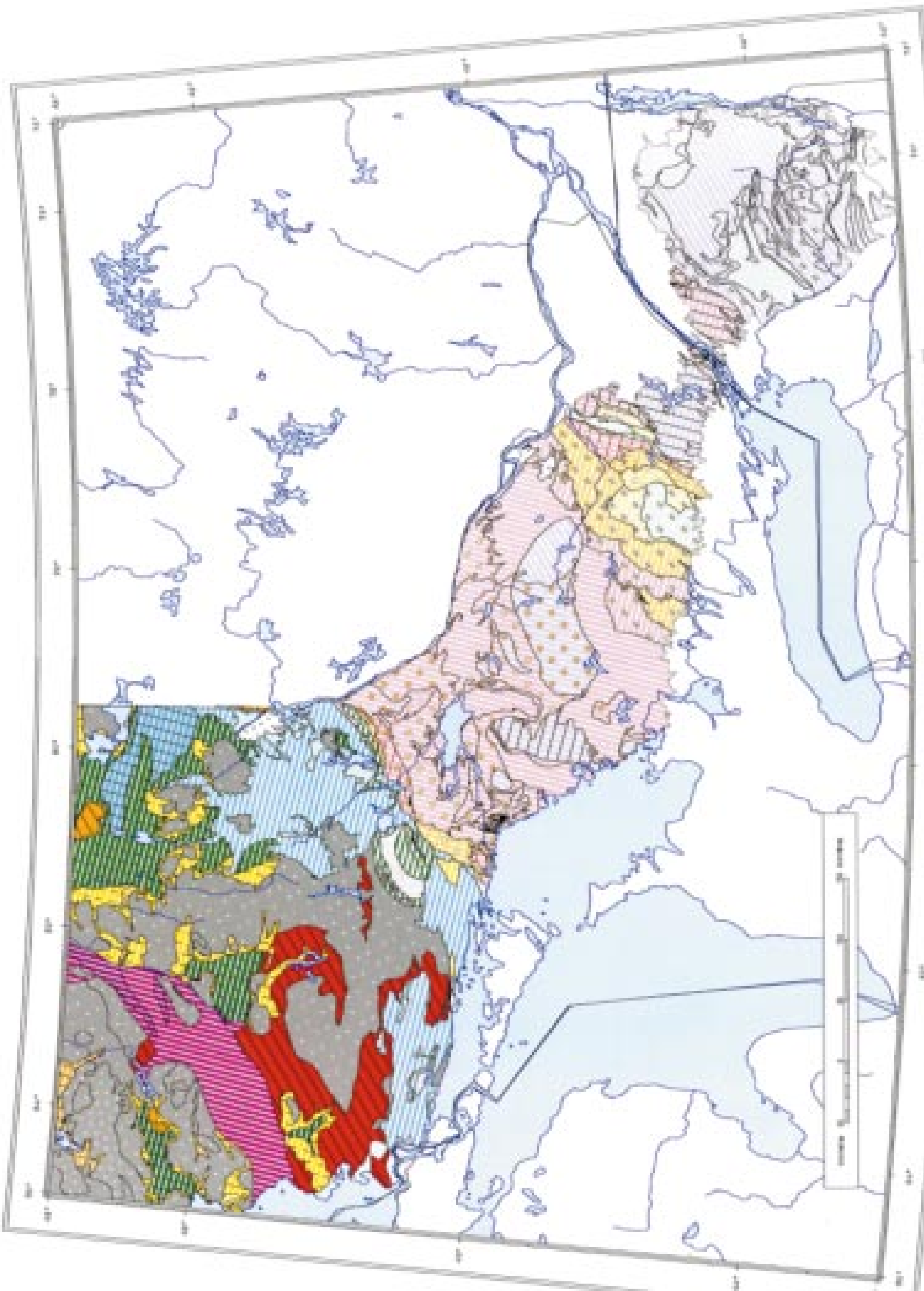
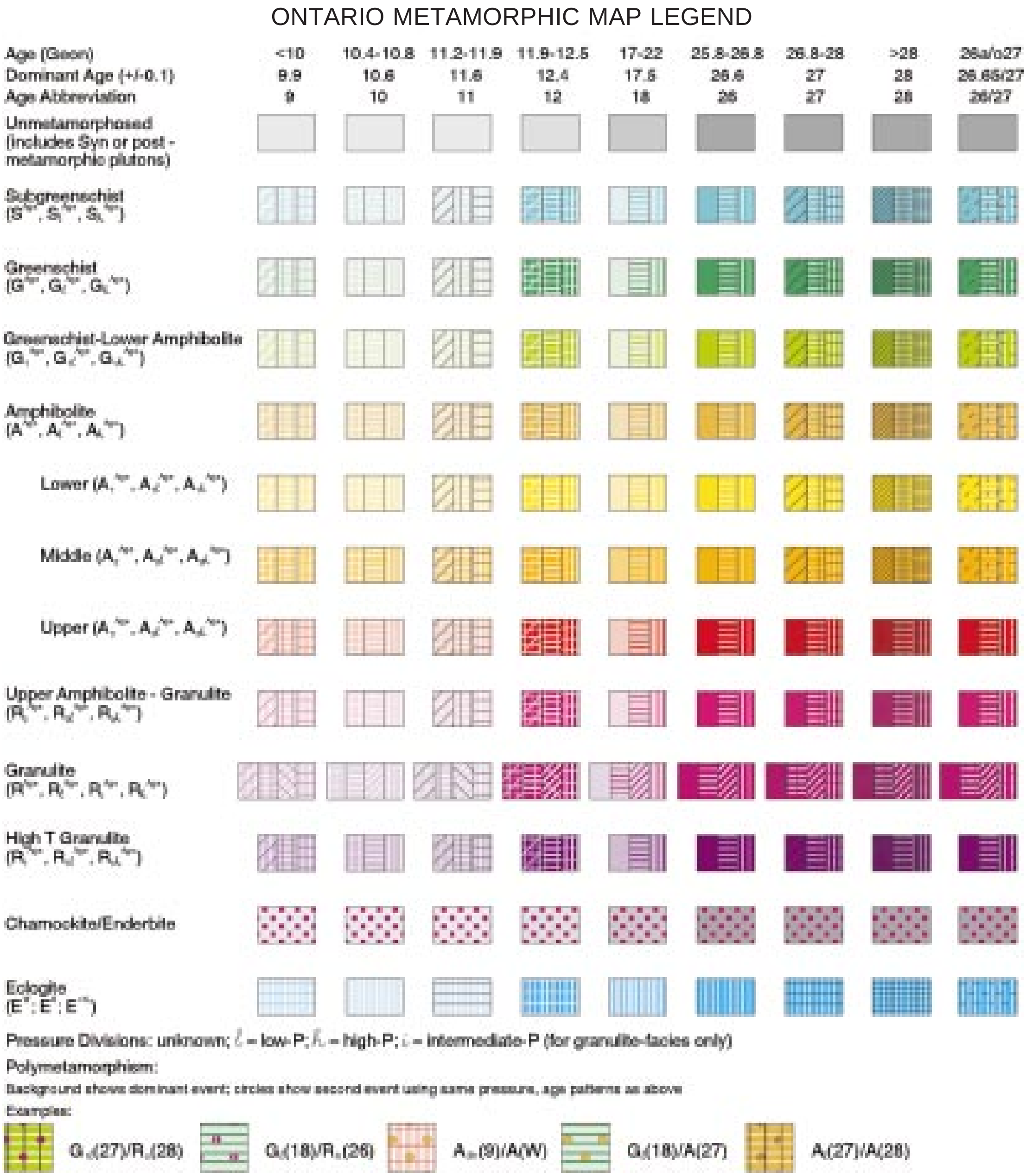


FIG. 2. Simplified version of the Southern and Grenville provinces, as illustrated on the Ontario portion of the Tectono-metamorphic Map of the Canadian Shield (this study). A brief explanation of the legend is given in part I (Easton 2000).



Animikie Basin

Within Ontario, the Animikie Group consists of the Gunflint and Rove formations (*e.g.*, Sutcliffe 1991, Morey 1983) (Figs. 1, 2). Deposition of the Animikie Group occurred prior to and during the Penokean Orogeny at ~1860 Ma. The best estimate of the age of deposition of these rocks comes from Fralick *et al.* (1998), who reported a U–Pb zircon age of 1878 ± 2 Ma from a lapilli tuff interbedded with sediments of the Gunflint Formation. Floran & Papike (1975) described primary greenalite, stilpnomelane, chamosite, and siderite from the Gunflint Formation, along with secondary magnetite, minnesotaite, ankerite and calcite. Fine-grained white mica is present in the Rove Formation (Geul 1970, 1973). All these assemblages indicate that the Animikie

Group underwent subgreenschist-facies metamorphism rather than diagenesis (Fraser *et al.* 1978). The age of this metamorphism is not well known, although in the United States, metamorphism and deformation of the Animikie Group occurred during the Penokean orogeny (*e.g.,* Morey 1983). ^{40}Ar–^{39}Ar ages from the Penokean orogen indicate that cooling occurred at ~1760 Ma (Holm *et al.* 1998). Metamorphism was pre-1537 Ma, the best estimate for the age of deposition of the overlying Sibley Group (Sutcliffe 1991). Rb–Sr ages from the Animikie Group in Ontario range from ~1400 to ~1660 Ma (Stille & Clauer 1986, Faure & Kovach 1969, Wanless & Loveridge 1978); samples were collected over a large area, however, and the ages may represent mixing lines. There is no indication that the Rb–Sr and K–Ar ages directly date the metamorphic minerals present in the Animikie Group. A whole-rock Nd–Sm age of 2076 Ma (Stille & Clauer 1986) likely reflects the age of the source region rather than the age of deposition or metamorphism.

Morey (1999) suggested that the Animikie Group in Minnesota was subjected to a regional alteration event related to the expulsion of brines during unroofing of the Penokean orogen. Distal effects of this fluid flow may be responsible for the young Rb–Sr and K–Ar ages from the Animikie Group. Without further study, it cannot be ascertained if the subgreenschist-facies assemblages reported from the Animikie Group in Ontario coincide with early amphibolite-facies metamorphism in the core of the Penokean orogen at 1870–1835 Ma, or if they coincide with late amphibolite-facies metamorphism and tectonic unroofing of the Penokean orogen at ~1760 Ma (*e.g.,* Holm *et al.* 1998). Because of this uncertainty, the Animikie Group is shown as affected by a Geon 18–16 subgreenschist-facies metamorphism on the metamorphic map (Fig. 2).

Where cut by Geon 11 Logan sills, a 2- to 15-cm-wide baked zone containing biotite has been reported within the Rove Formation (Geul 1973). Adjacent to the Geon 11 Duluth gabbro–anorthosite complex in northern Minnesota, a 2- to 8-km-wide pyroxene-hornfels facies contact metamorphic aureole is developed in rocks of the Animikie Group (*e.g.,* Morey 1978).

Sibley Group

The Sibley Group is younger than the Animikie Group but older than the Geon 11 Nipigon and Logan sills (*e.g.,* Sutcliffe 1991). A minimum age of deposition is given by a Rb–Sr age of 1340 ± 33 (Wanless & Loveridge 1978), with deposition most likely occurring at about 1537^{+10}_{-2} Ma, the U–Pb zircon age of a quartz–feldspar porphyry intrusion associated with volcanic fragmental rocks intercalated with the Sibley Group (Sutcliffe 1991). Metamorphic grade in the Sibley Group is lower than that in the Animikie Group. Fraser *et al.* (1978) showed the Sibley Group as subgreenschist facies; however, Franklin *et al.* (1980) described the presence of illite, smectite, chlorite, and authigenic microcline in Sibley Group shales. Whether this assemblage reflects diagenesis or regional metamorphism is problematic, and the Sibley Group can be shown as either unmetamorphosed or at subgreenschist facies. Following Fraser *et al.* (1978), it is shown as having undergone subgreenschist metamorphism between Geon 15 and 13. Adjacent to the Geon 11 Nipigon Sills, Rossport dolostones of the Sibley Group developed contact-metamorphic aureoles up to 10 m wide containing the assemblages Cal–Tr–Fo and Cal–Di–Fo (Sutcliffe 1991).

Keweenawan Supergroup

Rocks of the Keweenawan Supergroup were deposited in and marginal to the Midcontinent rift at approximately 1100 Ma (*e.g.,* Sutcliffe 1991) (Fig. 1), with the main period of volcanism in Ontario occurring between 1109 Ma and 1086 Ma (Davis & Sutcliffe 1985, Palmer & Davis 1987). Metamorphic grade within the Keweenawan Supergroup varies with stratigraphic position. In the Osler Lake Group on Black Bay Peninsula, the upper units are at zeolite-facies conditions, whereas the lower units are at the prehnite–pumpellyite facies (McIlwaine & Wallace 1976). The metamorphic minerals are commonly present as amygdule fillings within flows, especially in flow tops. In the zeolite-facies zone, the groundmass is commonly unaltered, other than showing devitrification. Within the prehnite–pumpellyite zone, however, especially in thin flows, the pyroxene is completely altered to chlorite, and plagioclase is converted to albite + white mica (McIlwaine & Wallace 1976). On Michipicoten Island, metamorphic grade appears to be lower, with an upper glassy zone (Annells 1974, Palmer *et al.* 1988) and a lower zone in the zeolite facies (Annells 1974). On the metamorphic map, the Keweenawan Supergroup is shown as having undergone a Geon 10–11 subgreenschist- facies metamorphism, but it should be noted that the intensity of metamorphism may increase with increasing stratigraphic thickness.

Recently, Puschner *et al.* (1999) suggested that the Portage Lake Volcanic Suite of the Keweenawan Supergroup in Michigan was subjected to two metamorphic events. The first is present throughout the entire stratigraphic section, and consists of a zeolite-dominated assemblage of laumontite – chlorite – corrensite ± wairakite ± Qtz ± Cal. The second event formed a Pmp–Ep assemblage, which is more common in the lower part of the sequence and in the vicinity of native copper deposits. Puschner *et al.* (1999) attributed these two events to a regional zeolite-facies burial metamorphism, followed by a localized, but where present, pervasive, hydrothermal event. Whether the same metamorphic history is present in the Osler Group, where similar assemblages have been reported (McIlwaine & Wallace 1976), remains to be determined, but could be significant in terms of identifying areas of copper mineralization.

The Southern Province

The geology of the Southern Province has been summarized by Bennett *et al.* (1991). Our knowledge of the metamorphic history of the Southern Province is based largely on the synthesis of Card (1978). The timing of metamorphism in the Southern Province is not well known, and has generally been assigned to the ~1870–1835 Ma Penokean Orogeny. Card (1978) recognized four metamorphic facies zones, namely, subgreenschist facies, lower- to middle-greenschist facies, middle- to upper-greenschist facies, and amphibolite facies. In the greenschist- and amphibolite-facies zones, metamorphism affected rocks of both the Huronian Supergroup and the Nipissing gabbro; consequently the timing of this event is younger than 2210 Ma, the age of the Nipissing gabbro (Noble & Lightfoot 1992). As elaborated upon below, whether regional metamorphism is older or younger than 1850 Ma Sudbury Igneous Complex (Krogh *et al.* 1984) has been a subject of considerable debate.

Consequently, a broad range in age, Geon 18–22, is shown on the metamorphic map (Fig. 2). Attempts to use ^{40}Ar–^{39}Ar methods to date metamorphism in low-grade Huronian metasedimentary units have yielded age groups at 1835–1810 Ma, 1765–1745 Ma, and 1550–1450 Ma (Hu *et al.* 1998). The first group is roughly "Penokean" in age; the others correspond roughly to the age of plutonic rocks present in the Southern and Grenville provinces. As noted by Fedo *et al.* (1997), however, much of the Huronian was subjected to regional K- and Na-metasomatism; consequently, the significance and robustness of these age groups, as well as previously determined Rb–Sr and K–Ar ages in the region (see summary in Easton 1986b), remain to be demonstrated.

Subgreenschist facies

Two broad areas of subgreenschist-facies metamorphism within the Huronian Supergroup were shown by Card (1978), one in the Sault Ste. Marie – Elliot Lake area, the other in the central and eastern Cobalt embayment (Figs. 1, 2). Both areas are characterized by a relatively thin sedimentary sequence (<6000 m), consisting mainly of rocks of the Cobalt Group. Metamorphic mineralogy in the eastern Cobalt embayment consists mainly of chlorite and muscovite porphyroblasts, with pyrophyllite also present in the central part of the embayment. The assemblage in the Cobalt embayment is characteristic of the subgreenschist facies (Frey 1987). Thomson (1966) reported the assemblage Prh–Pmp–Ab–Cal±Chl from the Henwood Nipissing gabbro, but on the basis of field relationships, attributed the assemblage to hydrothermal alteration.

In contrast, in the Sault Ste. Marie – Elliot Lake area, the dominant mineralogy is diaspore, kaolinite, illite–montmorillonite, stilpnomelane, chlorite, muscovite, and pyrophyllite; locally, andalusite has been reported (Chandler 1969, Wood 1973). Tectonic foliation is defined by muscovite and chlorite, and some Nipissing gabbro bodies are completely metamorphosed. The mineral assemblage in the Sault Ste. Marie – Elliot Lake area contains several mineralogical inconsistencies. The presence of diaspore (Chandler *et al.* 1969, Card 1978) is especially problematic. It can form as the result of hydrolysis of silicate minerals in a tropical climate, with the typical paragenesis being alumina + silica gel → diaspore → kaolinite, or it can form through hydrothermal alteration of aluminosilicate minerals. Alternatively, diaspore occurs in low-T – high-P environments, resulting from the breakdown of pyrophyllite. Minimum pressures for the development of diaspore in low-T – high-P regimes are >12 kbar (Theye *et al.* 1997), which is inconsistent with the indication of a low-P (<5 kbar) metamorphic regime in the Southern Province based on studies from the greenschist- and amphibolite-facies zones. The most logical explanation is that diaspore in the Huronian Supergroup is derived from either weathering or hydrothermal alteration. This interpretation also suggests that the mineral assemblages reported by Card (1978) from the Sault Ste. Marie – Elliot Lake are likely in disequilibrium, with phases such as diaspore, kaolinite, and illite–montmorillonite reflecting the weathering–alteration assemblage, and phases such as chlorite, muscovite, and pyrophyllite reflecting subgreenschist-facies regional metamorphism. Reports of andalusite and kyanite within metaquartzite in this zone (Church 1967, Chandler 1969) are also difficult to interpret; are they the result of prograde reactions, or the relics of an older event, now retrograded through alteration?

Whether the alteration that produced diaspore is pre- or post-regional metamorphism is unclear. Fedo *et al.* (1997) suggested that regional K-metasomatism occurred at 1728–1688 Ma, on the basis of Rb–Sr ages of Huronian paleosols (Roscoe *et al.* 1992). This event was followed closely by regional Na-metasomatism at 1700 ± 2 Ma, dated using U–Pb methods on monazite grains (Schandl *et al.* 1994). Fedo *et al.* (1997) also reported complete replacement of detrital grains of plagioclase in the Serpent Formation by albite; consequently, it is difficult to re-interpret previously published petrographic studies. Fedo *et al.* (1997) suggested fluid expulsion of brines due to uplift in the core of the Penokean orogen as a possible cause of this regional sodic alteration. The brines would require temperatures >110°C for the necessary albitization reactions to occur in both plagioclase and K-feldspar (Boles 1982); such temperatures can be reached at ~3 km depth in modern environments (*e.g.*, Boles 1982). Morey (1999) suggested a similar process for regional alteration in the Animikie basin at roughly the same time, *i.e.*, 1750 to 1700 Ma. Furthermore, Medaris *et al.* (1999) reported mineral assemblages similar to those found in the Huronian Supergroup from argillites in the

1760–1630 Ma Baraboo, Barron and Sioux quartzites, which also may be due to hydrothermal alteration, in their opinion.

This reconstruction leaves two possible interpretations of the mineral assemblages observed in the Sault Ste. Marie – Elliot Lake area: 1) Huronian sediments were intensely weathered after deposition, forming diaspore and kaolinite. This assemblage was subjected to subgreenschist-facies regional metamorphism, forming pyrophyllite, chlorite and muscovite. Regional metamorphism would need to be short-lived or patchy in order to preserve minerals formed during weathering. K- and Na-metasomatism may have followed locally. 2) Huronian sediments were metamorphosed to subgreenschist or higher metamorphic facies prior to ~1750 Ma, with the present assemblage of minerals representing a mixture of regional metamorphism and ~1700 Ma regional K- and Na-metasomatism. The evidence presented by Fedo *et al.* (1997), Morey (1999) and Medaris *et al.* (1999) suggests that the latter interpretation is the more likely. The Cobalt embayment may have been spared from this alteration by distance, as the alteration event seems to be centered near lakes Superior and Huron.

Gold mineralization has been reported in the Sudbury area in areas of intense Na-metasomatism (*e.g.*, Gates 1991). If the presence of diaspore is indicative of more intense alteration (*e.g.*, leaching of Na, Si and Fe), then similar mineralization might be expected throughout the Sault Ste. Marie – Elliot Lake area.

Greenschist and amphibolite facies

Higher-grade metamorphic rocks in the Southern Province occur mainly south of the Murray fault system in a set of elongate nodes, approximately east-trending (Card 1964, 1978), commonly coincident with major anticlinoria. This nodal style is similar to that recognized in rocks affected by the Penokean orogeny in northern Michigan (James 1955). More recent work has suggested that the nodes may be fault-bounded, rather than elliptical in form; however, limited surface exposure precludes rigorous definition of their geometry (Jackson 1998). In the Penokean orogen in Michigan and Wisconsin, Attoh & Klasner (1989) noted an association between the high-grade metamorphic nodes and Bouguer gravity lows. Modeling of the gravity field did not yield a unique solution; in some areas, the gravity low could be related to overthrusting, whereas in others, it was inferred to be due to an underlying migmatitic complex thought to have formed owing to crustal thinning during regional metamorphism. The high-grade metamorphic nodes in the Southern Province in Ontario are also spatially associated with Bouguer gravity lows; however, no attempt has been made to model the gravity field and relate it to potential sources of heat.

Card (1964) suggested that areas of higher metamorphic grade were associated with felsic intrusions, but, as noted by Jackson (1997), metamorphism occurred prior to intrusion of 1750 Ma plutons such as the Cutler and Eden Lake bodies; consequently, heating at 1750 Ma is unlikely. The 2220 Ma Nipissing gabbro is affected by regional metamorphism, making it unlikely that it was directly responsible for metamorphism, although Jackson (1997, 1998) suggested that crustal extension and mantle upwelling coincident with Nipissing magmatism may have supplied sufficient heat to produce the observed metamorphic pattern.

Kyanite and andalusite are stable in the Whitefish Falls area, well south of the Murray fault (Church 1967, Card 1978). In the Agnew Lake area, Fox (1971) reported andalusite in staurolite-grade rocks. In May Township, Jackson (1998) reported sillimanite in St–Bt rocks. The aluminosilicate assemblages suggest a maximum pressure of ~3.5 kbar. Thermobarometry on amphibolite-facies Grt–Bt–Ms rocks located south of the Murray fault in May Township, calculated by Jackson (1998) using TWEEQU (Berman 1991), gave pressures between 1.5 and 3 kbar, and temperatures of 500–560°C. Using TWEEQU on samples of metamorphosed Nipissing gabbro, Blonde (1996) estimated P–T conditions of 3.1 ± 2.5 kbar and 417 ± 60°C for greenschist-facies rocks north of the Murray fault, and 4.3 ± 0.8 kbar and 580 ± 60°C for amphibolite-facies rocks south of the Murray fault. These results are consistent with the observed aluminosilicate assemblages, and suggest low-pressure (2–4 kbar) metamorphism at a depth in the crust close to, or less than, the estimated thickness of the Huronian clastic wedge (Bennett *et al.* 1991). The lack of regional structural evidence for crustal thickening during the early fold-and-thrust event (Jackson 1997) makes orogenic thickening an unlikely mechanism for generating sufficient heat to produce amphibolite-facies metamorphism in the region.

Sudbury Igneous Complex and the Whitewater Group

Dressler (1984) summarized the distribution of shock-metamorphism features adjacent to the 1850 Ma Sudbury Igneous Complex (Krogh *et al.* 1984). Microscopic shock-induced textures have been destroyed in rocks of the South Range owing to later metamorphic recrystallization; such features are well preserved in the North Range, however. For example, planar lamellae in quartz grains occur up to a distance of 6–8 km from the contact (Dressler 1984). Dressler (1984) also documented a contact-metamorphic aureole in the North Range, consisting of a 100- to 200-m-wide pyroxene-hornfels zone, a 200-m-wide hornblende-hornfels zone, and a zone up to 1 km wide of plagioclase and quartz recrystallization. In the South Range, the contact-metamorphic aureole is ill defined, but relict green-brown hornblende cores in amphibolites of the Elsie Mountain Formation reported by Thomson *et al.* (1985) may be remnants of the hornblende-hornfels zone. James *et al.* (1992) reported greenschist alteration within rocks of the Levack gneiss complex in the footwall of the North Range discrete from the contact aureole of the Sudbury Igneous Complex; it is not clear however, if this alteration is related to the Sudbury Igneous Complex, or if it is associated with the low-grade regional metamorphism affecting the Southern Province.

North of the South Range shear zone, Fleet *et al.* (1987), Card (1978) and Rousell (1975) have all assigned rocks of the Sudbury Igneous Complex and the Whitewater Group to the subgreenschist or lower-greenschist facies, owing to the presence of chlorite, muscovite, and actinolite. In contrast, rocks south of the South Range shear zone show increasing metamorphic grade to the south and the east. The presence of biotite and spessartine in rocks of the Whitewater Group (Rousell 1975, Sadler 1958) attests to middle- to upper-greenschist-facies conditions (~400°C), and studies by Fleet *et al.* (1987) and Thomson *et al.* (1985) on rocks of the Sudbury Igneous Complex indicate that upper-greenschist to lower-amphibolite conditions were attained. Both Fleet *et al.* (1987) and Thomson *et al.* (1985) noted that the intensity of metamorphism in the mafic rocks depended on rock permeability. The metamorphic zones in the South Range mapped by Fleet *et al.* (1987) and Thomson *et al.* (1985) correspond closely to those mapped regionally by Card (1978), and were interpreted by Fleet *et al.* (1987) as indicating that regional metamorphism of the Southern Province occurred after emplacement of the Sudbury Igneous Complex, rather than before, as suggested by Card (1978). Timing of metamorphism is discussed further in the *Discussion* section.

Killarney belt

The Killarney belt consists of a northeastward-tapering wedge of predominantly plutonic rocks of ~1740 and ~1450 Ma age, stretching from Killarney to Coniston, that lie between rocks of the Huronian Supergroup and the Grenville Front. Metamorphosed equivalents of Killarney belt rocks can be traced into the Grenville Province (Bethune 1989, McGrath *et al.* 1988). Killarney plutons north of the Grenville Front have contact-metamorphosed adjacent Huronian Supergroup rocks up to 550 m distant from pluton margins (Card 1976). The contact aureoles contain chlorite, biotite and andalusite, characteristic of the hornblende-hornfels facies (Card 1976).

The Grenville Province

The metamorphism of the Grenville Province in Ontario has been previously summarized by Easton (1992). In this section, I update this previous summary, and include a summary of Grenvillian metamorphism in adjacent New York State. The organization of this section follows the nomenclature of Carr *et al.* (2000) (Fig. 3), who subdivided the Grenville Province into four main tectonic elements; namely, 1) Laurentia, 2)

TABLE 1. SUMMARY OF PRESSURE–TEMPERATURE ESTIMATES BASED ON THERMOBAROMETRY FOR GRENVILLIAN-AGE METAMORPHISM (1180–1000 Ma), ONTARIO AND NEW YORK STATE

Tectonic Domain	T (°C)	P (kbar)	Source
Grenville Front Tectonic Zone			
	no data	8–10	Anovitz & Essene (1990)
6.7 to 16.4 km SE of GF	720–760	6.1–8.4	Bethune & Davidson (1997)
1 to 4 km S of GF	660–685	7.4–8.1	Murphy (1999)
Parautochthonous domain (L2)			
Nipissing terrane	650–750	8–8.5	Anovitz & Essene (1990)
Tomiko terrane	<700	6–8	Anovitz & Essene (1990)
	725	8.4	Moore (1976)
Algonquin terrane	700–800	8–10	Culshaw *et al.* (1991)
Britt Domain			
Key Harbor gneiss	700–750	9–10	Corrigan (1990)
Nadeau Island gneiss	735	8.7	Culshaw *et al.* (1991)
	700–750	9–10.5	Anovitz & Essene (1990)
	700–730	6–11	Tuccillo *et al.* (1992)
Go Home Domain	710–720	8.8–10.3	Grant (1987, 1989)
	no data	9.5–10.5	Anovitz & Essene (1990)
Kiosk domain	750–825	9.5–10.5	Anovitz & Essene (1990)
Huntsville domain	700	9.2–9.9	Grant (1987, 1989)
	no data	8.5–10.5	Anovitz & Essene (1990)
	730–910	11.7–12.9	Timmermann (1998)
McClintock domain	670–715	7.9–8.8	Grant (1987, 1989)
	no data	8.5–10	Anovitz & Essene (1990)
recrystallized matrix	730–780	7.2	Timmermann (1998)
	800–820	9.6–9.8	Timmermann (1998)
McCraney domain	700–800	10–10.5	Anovitz & Essene (1990)
Novar domain	700–800	10.5	Anovitz & Essene (1990)
Opeongo domain	710–720	8.2–8.8	Grant (1987, 1989)
	700–750	8.25–9.25	Anovitz & Essene (1990)
Allochthonous domain (L3)			
Rosseau domain	700–800	8.5–10	Anovitz & Essene (1990)
Moon River domain	700–800	9.5–11	Anovitz & Essene (1990)
Seguin domain	700–800	11–12	Anovitz & Essene (1990)
Shawanaga domain			
Lighthouse gneiss	700–775	7	Wodicka *et al.* (2000)
Ojibway gneiss	685–700	7.5–8.5	Wodicka *et al.* (2000)
Muskoka domain			
migmatitic orthogneiss	720–780	7.5–10	Timmermann (1998)
metabasites	780–850	10.2–11.2	Timmermann (1998)
coronite	795	10.6	Timmermann (1998)
Parry Sound domain			
basal Parry Sound assemblage	700–750	8	Culshaw *et al.* (1991)
	650–725	8	Anovitz & Essene (1990)
~1120 Ma upper amphibolite event	615–700	5–7	Wodicka *et al.* (2000)
Parry Sound domain	690–710	9.1–9.8	Grant (1987, 1989)
	750–800	10–11	Anovitz & Essene (1990)
peak	885–975	12.5–13	Wodicka *et al.* (2000)
partial resetting	870–875	10–12	Wodicka *et al.* (2000)
later requilibration, ~1120 Ma?	780–870	7–8	Wodicka *et al.* (2000)
Composite Arc Belt			
Bancroft terrane	550–625	5.5–6.5	Anovitz & Essene (1990)
Elzevir terrane	375–425	3–4.5	Anovitz & Essene (1990)
Mazinaw terrane	450–550	4–5	Anovitz & Essene (1990)
Sharbot Lake terrane (east)	700–740	6.8–7.6	Buckley *et al.* (1997)
Frontenac–Adirondack Belt			
Frontenac terrane	600–700	4.25–6	Anovitz & Essene (1990)
Westport area	750	5.2	Lonker (1988)
	750–790	4.8–5.5	Buckley *et al.* (1997)
Adirondack lowlands	600–650	6–6.5	Anovitz & Essene (1990)
	600–725		Bohlen *et al.* (1985)
Adirondack highlands	670–780		Bohlen *et al.* (1985)
			Kitchen & Valley (1995)
	800–850	6.5–8	Spear & Markussen (1997)

GF: Grenville Front

a) Previous Nomenclature after Easton (1992)

b) This Study

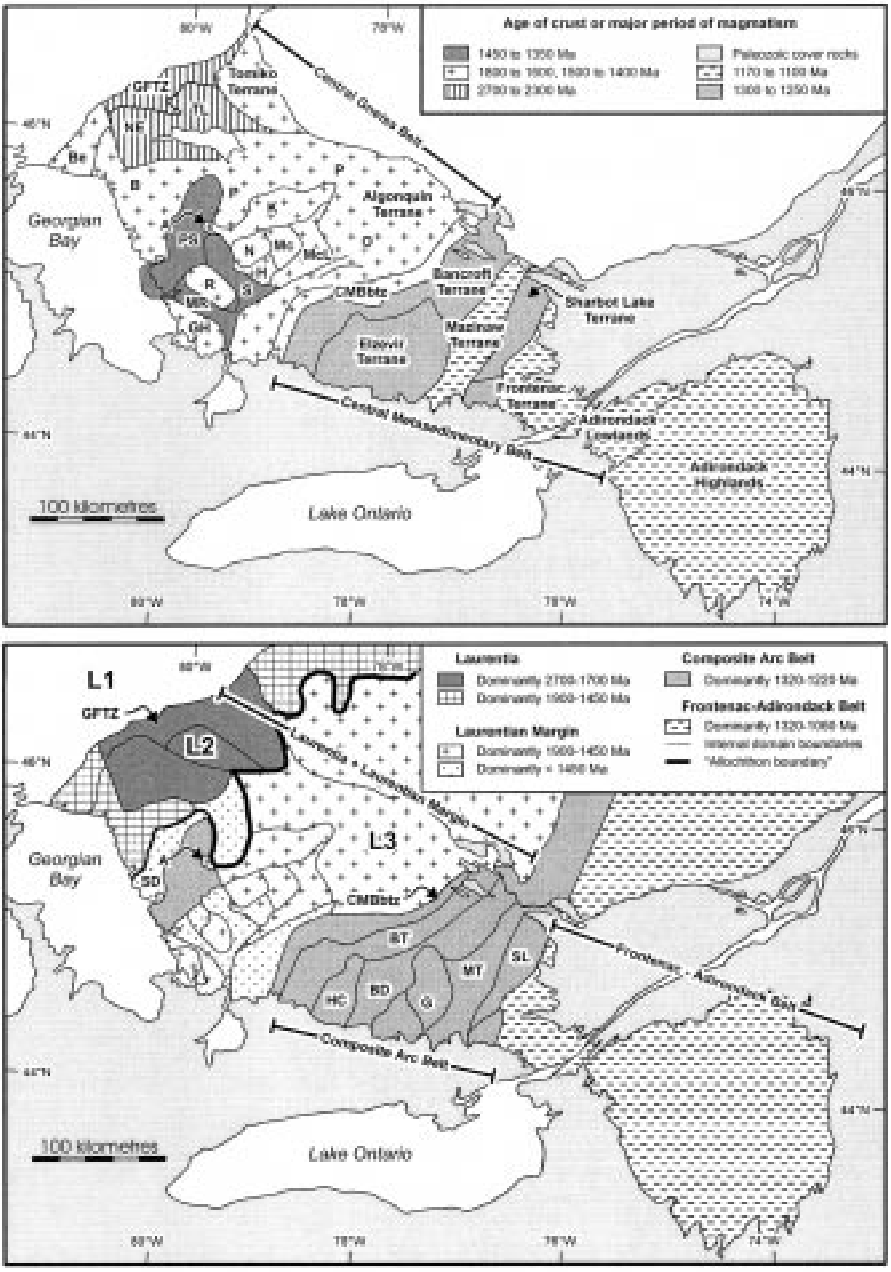


FIG. 3. Major divisions and structures of the southwestern Grenville Province, after Carr *et al.* (2000). a. Previous nomenclature showing lithotectonic terranes, domains and ages of crust, from Easton (1992). b. Nomenclature used in this paper, including the tripartite divisions:1) *Pre-Grenvillian Laurentia and its margin* (L2: Laurentian foreland northwest of the Grenville Front, L3: Archean crust with 1740 and 1450 Ma plutons, and L3: 1800–1680 Ma supracrustal rocks with ~1450 Ma continental arc granitic rocks), 2) *Composite Arc Belt*, and 3) *Frontenac–Adirondack Belt*. Abbreviations: A: Ahmic Domain, B: Britt Domain, BD: Belmont Domain, Be: Beverstone Domain (part of Killarney belt), BT: Bancroft Domain, CMBbtz: Central Metasedimentary Belt boundary thrust zone, G: Grimsthorpe Domain, GFTZ: Grenville Front Tectonic Zone, GH: Go Home Domain, H: Huntsville Domain, HC: Harvey Cardiff Arch, K: Kiosk Domain, Mc: McCraney Doamin, McL: McLintock Domain, MR: Moon River Domain, MT: Mazinaw Terane, N: Novar Domain, NE: Nepewassi Domain, O: Opeongo Domain, P: Powassan Domain, PS: Parry Sound Domain, R: Rosseau Domain, S: Seguin Domain, SD: Shawanaga Domain, SL: Sharbot Lake Domain, and TL: Tilden Lake Domain.

Laurentian margin, 3) the Composite Arc Belt, and 4) the Frontenac–Adirondack Belt. To use more familiar terminology, elements 1 and 2 comprise the Grenville Front tectonic zone, the Central Gneiss Belt, and the Central Metasedimentary Belt boundary zone; element 3 comprises the Central Metasedimentary Belt apart from Frontenac terrane, and element 4 comprises Frontenac terrane and the Adirondack Lowlands and Highlands, as illustrated in Figure 3. The term Grenvillian refers to events occurring between 1300 and 950 Ma, following the usage of Davidson (1998a).

As noted by Davidson (1998a), there is no regular arrangement of metamorphic facies parallel to the margins of the orogen. This in part reflects the polymetamorphic character of parts of the orogen, but in many areas it is due to tectonic juxtaposition that caused abrupt changes in metamorphic grade and, in some cases, age of metamorphism. Post-metamorphic-peak uplift also affected the pattern of metamorphism within the orogen. Nonetheless, in Ontario, some broad regional changes in the distribution and age of metamorphism can be related to large-scale crustal structure, as outlined below. Table 1 summarizes P–T conditions in the various tectonic divisions of the Grenville Province covered by this study.

Grenville Front Tectonic Zone

The Grenville Front Tectonic Zone (GFTZ) is an area ~30–50 km wide, located immediately south of the Grenville Front, which is characterized, in Ontario at least, by rock units that can be correlated with units north of the Grenville Front (*e.g.*, Davidson 1986, Easton 1992). Although rock units that can be equated with rocks of the adjacent Southern and Superior provinces and the Killarney belt can be identified within the GFTZ, to date, the grade and P–T conditions of any pre-Grenvillian metamorphism have yet to be documented in rocks from the GFTZ in Ontario. This is in contrast to Quebec, where Archean granulite-facies metamorphism overprinted by Grenvillian upper-amphibolite-facies metamorphism has been recognized in the GFTZ (*e.g.*, Indares & Martignole 1989, 1990a).

Within the GFTZ in the Killarney area, geochronological studies suggest an earlier high-grade metamorphic event, based on the presence of 1453 ± 7 Ma titanite in leucosome pods in metasedimentary gneisses (Krogh 1994), titanite ages from rocks along the north shore of Lake Huron that form an array with an upper intercept of 1454 ± 8 Ma (Haggart *et al.* 1993), and monazite ages of 1447–1440 Ma from paragneiss (Dudàs *et al.* 1994). These ages coincide with episodes of magmatism and granulite-facies metamorphism in the Britt domain to the southeast. Geochronology from the Sudbury area (Corfu & Easton 2000) revealed metamorphic events at ~1720 Ma and ~1450 Ma on the basis of titanite and zircon ages. Furthermore, in the Killarney and Sudbury areas, the diabase dikes of the ~1240 Ma Sudbury swarm cut earlier gneissic fabrics in the country rocks, suggesting the presence of a pre-1240 Ma leucosome-forming event. In detailed metamorphic studies in the Sudbury area, Murphy (1999) failed to find mineralogical relics of older metamorphic events. This may simply attest to the intensity of Grenvillian metamorphism in obliterating any pre-existing metamorphic mineralogy.

P–T conditions of the upper-amphibolite facies Grenvillian metamorphism are estimated at 660–685°C and 7.4–8.1 kbar within 1 to 4 km southeast of the Grenville Front (Murphy 1999) and 720–760°C and 6.1–8.4 kbar at distances of 6–17 km southeast of the Grenville Front (Bethune & Davidson (1997). Bethune & Davidson (1997) proposed a clockwise P–T–t path, with nearly isobaric cooling following peak metamorphism. The timing of Grenvillian metamorphism within the GFTZ is generally younger than that observed throughout the Central Gneiss Belt, much of the Composite Arc Belt, and the Frontenac–Adirondack Belt, with metamorphic zircon and lower intercept ages clustering in the ~990 Ma range (*e.g.*, Krogh 1994, Haggart *et al.* 1993, Corfu & Easton 2000). Monazite and titanite ages from the GFTZ also fall in the 990–980 Ma range. Cooling history curves are difficult to construct for the GFTZ, as excess Ar is common in minerals found on both sides of the Grenville Front (*e.g.*, Wanless *et al.* 1970, Haggart *et al.* 1993, Smith *et al.* 1994, Reynolds *et al.* 1995). Cooling history curves constructed using a combination of U–Pb and Ar–Ar data (Haggart *et al.* 1993), or U–Pb data alone (Corfu & Easton 2000, Murphy 1999) suggest rapid cooling, on the order of 7–10°C/m.y. from 990 to 960 Ma, with slower cooling of 2–4°C/m.y. in the interval 960–940 Ma.

Laurentia and Laurentian margin (*Central Gneiss Belt*)

Although Laurentia and the Laurentian margin (Central Gneiss Belt) can be subdivided on the basis of geology, structure and geophysical characteristics into several domains and terranes (*e.g.,* Culshaw *et al.* 1983, Davidson 1984, Easton 1992), it is more useful for the purpose of this paper to subdivide this region based on affinity to the North American craton (Carr *et al.* 2000) (see Fig. 3). The three resulting divisions are: 1) a parautochthonous domain, consisting of rocks that were likely part of Laurentia, and which contains remnants of Sudbury diabase dikes (L2 in Fig. 3b), 2) an allochthonous domain, consisting of rocks that may have formed on the margin of Laurentia, and which locally contains ~1170 Ma mafic intrusions and fragments of eclogites (*e.g.*, Ketchum & Davidson 2000) (L3 in Fig. 3b), and 3) an allochthonous domain in the Parry Sound region that was probably exotic (*e.g.*, Wodicka *et al.* 1996, Culshaw *et al.* 1997, Carr *et al.* 2000). Metamorphic history varies among each of these segments, as outlined below and in Table 1. Apart from the presence of ~1160 Ma metagabbros in L3, and late dikes of granitic pegmatite, there are no major plutonic units of Grenvillian age (1300–950 Ma) in the Central Gneiss Belt.

Laurentia and Laurentian margin (*Central Gneiss Belt*) – *Parautochthonous domain* (*L2*)

Because of its close affinities to the North American craton, and the presence of Archean, Paleoproterozoic and Mesoproterozoic rocks, it is not surprising that this region locally preserves evidence of pre-Grenvillian metamorphic history. In the Key Harbour area (Britt domain), Corrigan *et al.* (1994) described at least three metamorphic events: 1) an upper-amphibolite-facies event marked by migmatitic gneisses cross-cut by the ~1694 Ma Key Harbour leucogranite, 2) an upper-amphibolite- to granulite-facies event affecting the Key Harbour gneiss and the 1694 Ma leucogranite, but not the ~1450 Ma plutons, and 3) a granulite- to upper-amphibolite-facies Grenvillian event that affected all rocks in the area, including the ~1240 Ma Sudbury diabase dikes. Corrigan *et al.* (1994) suggested that during the ~1070 Ma Grenvillian event, conditions in the granulite facies may have been attained initially, re-equilibrating to the upper amphibolite facies at lower P–T and higher P(H_2O) conditions. Peak conditions of Grenvillian metamorphism are estimated at 860°C and 12 kbar (Culshaw *et al.* 1997). P–T–t paths from the Britt domain and the transition zone into the GFTZ both suggest near-isothermal decompression over the range 14–6 kbar and 860–680°C (Jamieson *et al.* 1995). The lack of eclogite development under these conditions of pressure is attributed to limited residence-time at peak P and T (Jamieson *et al.* 1995).

In the Pointe-au-Baril area (Britt domain), Ketchum *et al.* (1994) recognized an area of 1452 ± 2 Ma granulite-facies metamorphism that has been overprinted by the widespread 1070–1035 Ma upper amphibolite facies typical of the Grenvillian event in the L2 region (Culshaw *et al.* 1997, Carr *et al.* 2000). In the McClintock domain, north of Dorset, Timmermann (1998) reported evidence for ~1450 Ma metamorphism. In both areas, this ~1450 Ma metamorphism is associated with an influx of a suite of similar-aged felsic plutons across the L2 region. It is probable that ~1450 Ma metamorphism is preserved elsewhere in the L2 region, which is generally poorly studied apart from the Georgian Bay area. On the metamorphic map (Fig. 2), this 1450 Ma event is shown only where it is known to occur.

P–T conditions of the ~1450 Ma granulite-facies event have been estimated at 625–700°C and 7.2–8.4 kbar, with the low temperatures possibly reflecting low activity of H_2O during metamorphism (Ketchum *et al.* 1994). In contrast, conditions of Grenvillian metamorphism, 720–775°C and 10.8–11.5 kbar, were attained (Ketchum *et al.* 1994). Ketchum *et al.* (1994) and Tuccillo *et al.* (1990) both suggested a single isothermal decompression P–T–t path, from ~11 to 6 kbar during Grenvillian orogenesis. In contrast, Tuccillo *et al.* (1992) combined the results from both metamorphic episodes and proposed a clockwise P–T–t path.

Apart from the 1450 Ma granulites, the only other area of granulite facies rocks in the L2 region is a broad band extending southwest of Port Loring (Davidson *et al.* 1982). The age of this granulite-facies event is unknown; on the map (Fig. 2) it is portrayed as part of the 1070–1040 Ma Grenvillian event. Limited data for Grenvillian upper-amphibolite-facies metamorphism in the L2 domain, mainly from the Georgian Bay and Port Loring areas, suggest P–T conditions of 700–800°C and 9–11 kbar (Anovitz & Essene 1990) (Table 1).

Laurentia and Laurentian Margin (*Central Gneiss Belt*) – *Allochthonous domain* (*L3*)

The oldest rocks in this domain are ~1450–1400 Ma (Ketchum & Krogh 1997) eclogitic rocks and felsic plutons. Vast tracts of migmatitic gneiss, including some of supracrustal origin, remain undated, but likely formed in the interval 1450–1300 Ma (Culshaw *et al.* 1997, Carr *et al.* 2000, and references therein). This domain contains pods of coronitic metagabbro, emplaced at ~1160 Ma (van Breemen & Davidson 1990).

Grt–Cpx metabasites interpreted as retrograded eclogite (Davidson 1991) occur in two areas, as small pods along the L2/L3 boundary in the Georgian Bay to Emsdale region, and northwest of Mattawa. Pods of retrograded eclogite are typically associated with anorthosite blocks, particularly in the L2/L3 bounding shear zone, consistent with derivation of the eclogites from mafic intrusions (*e.g.,* Davidson 1991, Culshaw *et al.* 1997). Grant (1987, 1989) and Davidson (1991) studied the eclogitic rocks in the Georgian Bay to Huntsville area in detail. The Mattawa area eclogites have not been studied, and their extent is known mainly because of interest in exploiting these rocks as a source of garnet for use in the abrasives industry. Grant (1989) estimated P–T conditions of ~14 to 16 kbar and temperatures of 775–875°C, and suggested that the emplacement of the eclogitic and related sapphirine-bearing rocks involved rapid transport from depths of 55 km to higher levels in the crust (25–30 km), and tectonic emplacement into the surrounding gneisses. Davidson (1991), on the other hand, suggested that the surrounding quartzofeldspathic gneisses were subjected to the same high-pressure conditions, but that they re-equilibrated completely to lower-pressure conditions as a result of continuous ductile strain. If Davidson (1991) is correct, then the area subjected to high-pressure metamorphism was more extensive than is currently shown on the metamorphic map (Fig. 2). The eclogites contain metamorphic zircon that formed between 1085 and 1095 Ma (Ketchum & Krogh 1997), roughly consistent with the older ages of metamorphic zircon from this domain, which are in the range 1060–1080 Ma (*e.g.,* Timmermann *et al.* 1997, Bussy *et al.* 1995). Coronitic metagabbros yield slightly younger ages of metamorphic zircon, ~1050 Ma (van Breemen & Davidson 1990, Heaman & LeCheminant 1993), attributed to slow diffusion (T.E. Krogh, pers. commun., 2000).

In a zone extending up to 75 km inboard of the boundary with the Composite Arc Belt, extensive tracts of granulite facies occur in the L3 domain, as shown on the metamorphic map (Fig. 2). All rocks in the granulite facies exhibit retrogression at local and regional scales, especially where they are cut by younger dikes and extensional shear zones (*e.g.,* Davidson 1991). Estimated conditions for the granulite and upper amphibolite facies from L3 (see Table 1) are generally in the range 670–720°C and 8–10 kbar, possibly some 50–80°C and 1–2 kbar lower than conditions in the southern L2 region. P–T–t paths from the Georgian Bay area show decompression accompanied by some cooling (from 830°C, 11 kbar to 550°C, 5 kbar: Wodicka *et al.* 2000).

The age of the granulite-facies metamorphism within L3 is unknown, and it is possible that two ages of granulite-facies metamorphism are present: 1) an earlier event that is pre-emplacement of the ~1160 Ma coronitic metagabbros; this episode would be consistent with the expectation that rocks in the footwall of L3 would exhibit higher P–T conditions owing to crustal thickening during thrusting; 2) a younger event (1080–1050 Ma) resulting from crustal thickening in response to emplacement of the Composite Arc Belt over Laurentia and the Laurentian margin during the Grenville orogeny. This episode would be consistent with zircon ages of ~1080–1060 Ma (*e.g.,* van Breemen *et al.* 1986, Timmermann *et al.* 1997) and the occurrence of late syn- to post-tectonic, orthopyroxene-bearing pegmatite dikes in the Dorset and Barry's Bay area. Clearly, more detailed work is needed to resolve the P–T history of the L3 domain.

A distinguishing feature of L3 domain is the prodigious development of migmatite throughout the domain, particularly near the boundary with the parautochthonous domain (L2) (*e.g.,* Culshaw *et al.* 1997, Carr *et al*. 2000, and references therein). The cause of migmatization is problematic, although once formed, it facilitated thrust transport (*e.g.,* Culshaw *et al.* 1997). Migmatites in the Muskoka domain formed between 1080 and 1065 Ma (Timmermann *et al.* 1997, Timmermann 1998). Peak ages of metamorphism from L3 are generally in the range ~1080–1060 Ma (*e.g.,* Timmermann *et al.* 1997, Bussy *et al.* 1995), the same age as late syntectonic pegmatites in the region (*e.g.,* Nadeau & van Breemen 1990, van Breemen & Hanmer 1986) and the development of zircon in strain shadows during regional deformation (*e.g.,* Bussy *et al.* 1995).

Laurentia and Laurentian Margin (*Central Gneiss Belt*) *– Parry Sound allochthonous domain*

The Parry Sound domain is composed mainly of plutonic rocks of mafic to intermediate composition and supracrustal rocks that were metamorphosed to the granulite facies at ~1160 Ma (van Breemen *et al.* 1986, Tuccillo *et al.* 1992, Wodicka *et al.* 1996), distinct from the 1070–1040 Ma age range typical of the L2 and L3 domains. Estimates of peak conditions of metamorphism from the interior of Parry Sound domain are 885–975°C and 12–13 kbar (Wodicka *et al.* 2000), although somewhat lower estimates (800°C, 10–11 kbar) have been previously reported (Table 1), with pelitic rocks containing the assemblage Grt–Sil–Opx–Kfs–Qtz–Spl–Bt. P–T–t paths show decompression accompanied by some cooling (*e.g.,* Wodicka *et al.* 2000, Anovitz & Essene 1990), similar to the path reported from the adjacent Shawanaga domain (L3) and from the Britt domain (L2). In the southern Parry Sound domain, metamorphic zircon, monazite, and titanite ages exhibit a wide range, from as old ~1120 Ma (Ketchum & Krogh 1997, Krogh *et al.* 1999), to ~1050 Ma. Krogh *et al.* (1999) suggested that the spread in ages reflects the fact that this segment of the crust was hot over most of the interval 1160 to 1060 Ma, with zircon forming as a result of a variety of factors related to regional fluid flow, ambient conditions of temperature, *etc*.

Composite Arc Belt

Metamorphic conditions within the various geological domains of the Composite Arc Belt are summarized in Table 1. As illustrated in Figure 4a, the previous view of this area was that this region experienced a single metamorphic episode, with the lowest metamorphic grade occurring in the Madoc area, and with metamorphic grade increasing to the east, north and west. As shown in Figure 4b, the pattern of metamorphism is more complex, with major structures juxtaposing rocks of different metamorphic grade and age (*e.g.,* Mazinaw and Sharbot Lake domains), or juxtaposing domains with a complex metamorphic history against domains displaying an apparently simple metamorphic history (*e.g.,* Mazinaw and Grimsthorpe domains).

On a regional scale, there are at least two major metamorphic episodes. The first occurs at ~1250–1240 Ma, and is associated with accretion of various supracrustal assemblages and the influx of a suite of granitic and gabbroic intrusions. Many of these plutons produced contact metamorphic aureoles, and it is possible that a regional metamorphic event may have occurred in this time period related to the large volume of magma emplaced into the crust. The best documented contact-aureole of this event occurs adjacent to the Tudor gabbro, where calcite–graphite thermometry in marbles adjacent to the body yields temperatures of 450–500°C two km from the intrusion, and 700–750°C at the contact (Dunn & Valley 1992), consistent with mapped isograds around the intrusion (Allen 1976, LeAnderson 1978). In contrast, the calcite–dolomite thermometer was reset during regional metamorphism, yielding temperatures of ~500°C even at the contact

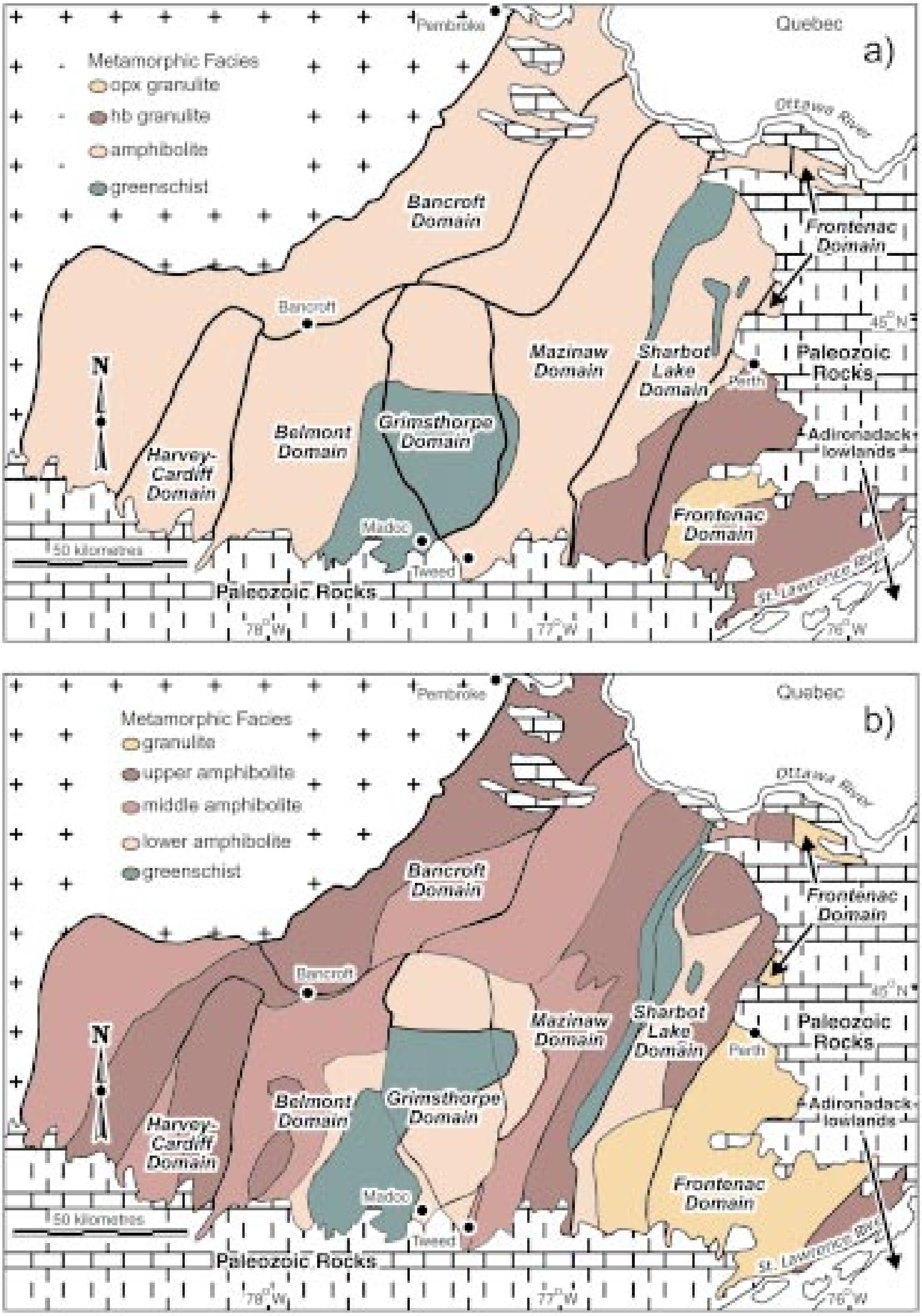


FIG. 4. Metamorphic grade within the Composite Arc Belt, Grenville Province, Ontario. a. Map showing distribution of metamorphic facies prior to 1980. b. Map showing the distribution of metamorphic facies as of 1999. Note that the age of metamorphism is not the same across the Belt, *e.g.*, amphibolite-facies metamorphism in the Mazinaw domain is some 50–60 Ma younger than amphibolite-facies metamorphism in Belmont domain.

(Allen 1976, Dunn & Valley 1992). Evidence for an early regional metamorphic event is found chiefly in Belmont and Sharbot Lake domains, where 1245–1229 Ma gabbro plutons cut metamorphic layering present within marbles. This finding suggests development of a regional fabric (and recrystallization) in the marbles prior to, or coincident, with pluton emplacement.

The second event occurred at 1070–1060 Ma, and is best documented within the Belmont domain, based on limited data on titanite from felsic plutonic rocks (*e.g.,* Davis & Bartlett 1988). This information is also consistent with hornblende $^{40}Ar–^{39}Ar$ ages of ~1050–1020 Ma from the Belmont domain (*e.g.,* Cosca 1989). The timing of the 1070–1060 Ma metamorphism in the Composite Arc Belt is coincident with regional metamorphism and deformation within the Parautochthonous (L2) and Allochthonous (L3) domains, and in the Adirondack Highlands. Grade ranges from the greenschist facies in the Madoc area to the upper amphibolite facies adjacent the northern margin of the belt (Fig. 4b). Regionally, a suite of potassic plutons ranging in age from 1085 to 1066 Ma was emplaced into the Composite Arc Belt during this time; the volume of magma associated with this episode of plutonism is unlikely to have served as a major heat-source, however. Further, many of these plutons possess contact-metamorphic aureoles, and are generally unaffected by the regional metamorphic event (*e.g.,* Corriveau 1990).

A slightly different metamorphic history is present in the two westernmost domains, namely the Sharbot Lake and Mazinaw domains (Figs. 2, 4b). In western Sharbot Lake domain, greenschist- to amphibolite-facies metamorphism occurred at ~1200 Ma, on the basis of limited U–Pb data on titanite (Corfu & Easton 1997). In contrast, in the eastern and southern Sharbot Lake domain, the grade of metamorphism ranges from the middle to upper amphibolite facies, and to the granulite facies if the Hinchinbrooke pluton is included within the Sharbot Lake domain, as has been suggested by Easton (1999a). In the Wolf Grove area of the Sharbot Lake domain, Corfu & Easton (1997) dated this metamorphism at ~1168 Ma; Buckley *et al.* (1997) estimated the P–T conditions in the upper-amphibolite-facies rocks at 650–700°C and 7–8 kbar. Easton & Davidson (1997) and Easton (1999a) suggested that the southern and eastern Sharbot Lake domain was overprinted by the ~1168 Ma high-grade metamorphism present within Frontenac terrane, as a consequence of the overthrusting of the Frontenac terrane onto the Sharbot Lake domain along the Maberly shear zone. Evidence for the 1070 Ma metamorphic event seems to be lacking within Sharbot Lake domain, as well as in the adjacent Frontenac domain (Fig. 4b). This lack is consistent with the view that Sharbot Lake and the Frontenac–Adirondack belt became linked at ~1160 Ma (*e.g.,* Carr *et al.* 2000) and were at high levels in the crust at ~1070 Ma.

The east-central Mazinaw domain is in the greenschist facies; grade increases northwestward to the upper amphibolite facies (Fig. 4b). Metasedimentary rocks of the Flinton Group, which lie unconformably on supracrustal rocks of the Grenville Supergroup, and which contain detrital zircon as young as 1155 Ma (Sager-Kinsman & Parrish 1993), allow mapping of a number of regional metamorphic isograds within Mazinaw domain (*e.g.,* Easton 1996). The Mazinaw domain may have experienced three metamorphic episodes, the aforementioned ~1240 Ma contact–regional metamorphic event, a ~1070 Ma regional metamorphic event, and a younger ~1020 Ma episode of regional metamorphism. Evidence for the 1070 Ma event is based on the presence of zircon ages of 1080 Ma from pegmatite veins cutting tonalite gneisses in the northern part of the domain, as well as a titanite age of 1052 Ma from calc-silicate rocks (Corfu & Easton 1995). The 1070 Ma event was largely obliterated by the younger event, which formed the observed greenschist- to upper-amphibolite-facies zonation in the domain, and which occurred between 1036 and 1020 Ma (Corfu & Easton 1995). The timing of this event is constrained by metamorphic zircon from upper-amphibolite-facies metavolcanic rocks and metamorphic zircon and monazite from upper-amphibolite-facies metapelites of the Flinton Group. Hornblende $^{40}Ar–^{39}Ar$ ages from the Mazinaw domain, ~940 Ma (Cosca 1989), are roughly 60–80 Ma younger than most other parts of the Composite Arc Belt (Fig. 5), but comparable to ages from the Adirondack Highlands.

Frontenac–Adirondack Belt – Frontenac domain

The Frontenac domain is characterized by a low-pressure granulite-facies metamorphism that occurred at, or prior to, 1168–1162 Ma, the age of metamorphic zircon and titanite from the domain (Corfu & Easton 1997, Mezger *et al.* 1993). Unlike many other areas of the Grenville Province in Ontario, this metamorphism is associated with emplacement of large volume of felsic and subordinate mafic magmas of the Frontenac Suite at 1180–1155 Ma. Post-metamorphic cooling appears to be have been rapid, with $^{40}Ar–^{39}Ar$ hornblende ages between 1125 and 1110 Ma (Cosca 1989) (Fig. 5). Wynne-Edwards (1967) showed a symmetrical distribution of metamorphic grade associated with this event, with conditions peaking south of Westport, and decreasing to the north and south. Carmichael *et al.* (1978) and Lonker (1980), however, suggested that the highest metamorphic grade is found north of Gananoque, and that no culmination is present near Westport. This difference in opinion reflects the fact that many of the granulites contain disequilibrium textures, making determination of peak conditions difficult. P–T determinations by Lonker (1988) and Buckley *et al.* (1997) suggest conditions of 700–750°C and pressures of 4.5–6 kbar (Table 1).

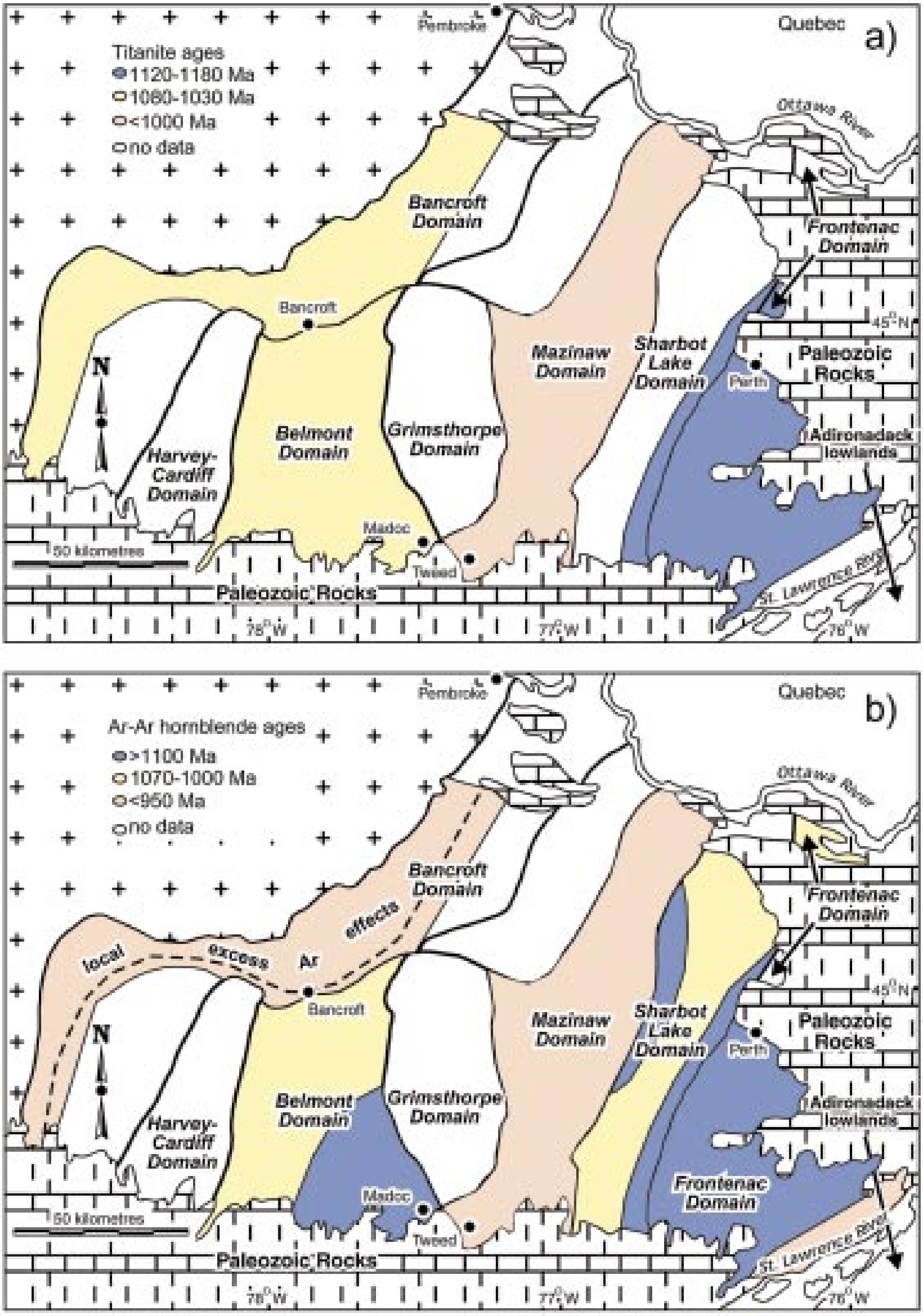


FIG. 5. Cooling history of the Composite Arc Belt and the Frontenac–Adirondack belt, Grenville Province, Ontario. a. Ages based on metamorphic titanite. b. ^{40}Ar–^{39}Ar hornblende ages (data from Easton 1986a and subsequent updates).

Frontenac–Adirondack Belt – Adirondack Lowlands

This region contains many of the same rock units as the Frontenac domain, but has been subjected to middle- to upper-amphibolite-facies metamorphism. Peak P–T conditions summarized in Bohlen *et al.* (1985) are 600–725°C and 6–6.5 kbar; the resultant P–T–t path indicates near-isobaric cooling. Metamorphism was synchronous with that in Frontenac domain, at ~1168 Ma, on the basis of U–Pb ages of titanite and garnet from metapelites and marbles in the lowlands (Mezger *et al.* 1991, 1992). In contrast to the Canadian side, hornblende ^{40}Ar–^{39}Ar ages in the lowlands are ~40–60 Ma younger (1020–1080 Ma: Magloughlin *et al.* 1997). The younger ^{40}Ar–^{39}Ar hornblende ages may reflect heating above 500°C during the 1080–1070 Ma granulite-facies metamorphism present in the Adirondack Highlands to the south. The presence of an Opx-in isograd (Hoffman 1982, De Waard 1965) in the lowlands parallel to, but west of the Carthage–Colton mylonite zone, which is generally regarded as the Lowlands–Highlands boundary, is consistent with this interpretation. This reconstruction assumes, however, that the Opx-in isograd is ~1080 Ma in age, and the isograd is not related to an older metamorphic event with its current location reflecting uplift along the Carthage–Colton mylonite zone. In contrast to the Opx-in isograd, isotherms based on oxide and feldspar thermometry in the Lowlands (Bohlen *et al.* 1985) do not show any systematic pattern with respect to the Lowlands–Highlands boundary. These isotherms, however, record temperatures near or above titanite-closure temperatures, and may have been unaffected by younger thermal events.

Frontenac–Adirondack Belt – Adirondack Highlands

Most U–Pb age determinations from the Adirondack Highlands are from plutonic rocks; consequently, the history of plutonism has greatly influenced interpretation of the tectonic and metamorphic history of the Highlands. Three main plutonic suites have been identified in the Adirondack Highlands: the ~1150–1125 Ma anorthosite – mangerite – charnockite – granite (AMCG) suite, the ~1100–1090 Ma Hawkeye granitic suite, and the ~1070–1040 Ma Lyon Mt. suite (*e.g.*, McLelland *et al.* 1996). In the southern Highlands, older rocks are reported, including a ~1300–1330 Ma calc-alkaline suite (McLelland & Chiarenzelli 1990, McLelland *et al.* 1996) and a ~1240 Ma charnockitic suite (Chiarenzelli & McLelland 1991, McLelland *et al.* 1996). Supracrustal rocks are older than AMCG magmatism, but whether or not more than one supracrustal package is present in the Highlands is, as yet, undetermined.

Although the Adirondack Highlands are well studied, there is no consensus regarding metamorphic history. Important areas of disagreement include: (1) the presence, P–T conditions, and metamorphic type (regional, contact, *etc.*) of a pre-1130 Ma metamorphic event, (2) the depth of emplacement of the massifs of anorthosite and their role as a heat source for metamorphism, (3) the nature of the P–T–t path for the Highlands, with clockwise, anticlockwise, and isobaric paths all proposed, and (4) the presence of systematic trends in metamorphic conditions across the central Highlands, and if so, their significance.

The Adirondack Highlands are in the granulite facies, with oxide, feldspar and calcite–graphite thermometry suggesting concentric isotherms, centered immediately west of Lake Placid over the western lobe of the Marcy anorthosite (*e.g.*, Bohlen *et al.* 1985, Kitchen & Valley 1995). The timing of this metamorphism is not well constrained, but likely occurred after 1080 Ma, the age of a mangerite unit in the northwest Highlands (McLelland *et al.* 1996). Metapelites from the Highlands yield garnet and monazite ages of 1064 and 1033 Ma, respectively (Mezger *et al.* 1991), and Florence *et al.* (1995) reported metamorphic U–Pb monazite ages of 1032–1049 Ma from the Port Leyden area in the southwestern Adirondacks, suggesting that the metamorphic peak occurred prior to ~1060 Ma. Using two-pyroxene thermometry, Spear & Markussen (1997) did not observe any systematic trends in temperature across the Highlands, and suggested that the pattern of concentric isotherms reflects some aspect of the thermal maximum that developed prior to the appearance of garnet, which crystallized late in the cooling history.

A pre-1080 Ma upper amphibolite- to granulite-facies metamorphic event in the southern Adirondack Highlands is suspected on the basis of the state of deformation of the older plutonic suites present there. In the northern and central Adirondack Highlands, a pre-1140 Ma regional metamorphic event of upper amphibolite to granulite facies also is suspected because of the presence of garnet–sillimanite gneiss xenoliths cut by 1144 ± 7 Ma gabbroic rocks (McLelland & Chiarenzelli 1990, McLelland *et al.* 1996). Anatexis in metapelites in the Port Leyden area is older than 1130 Ma (Florence *et al.* 1995), also suggesting an earlier high-grade event.

In contrast, Alcock & Muller (1999) suggested that migmatitic mangeritic gneisses of the New Russia complex, cut by the ~1130 Ma Marcy anorthosite, represent the contact-metamorphic aureole of the intrusion. Contact metamorphism could account for the observations made by McLelland *et al.* (1996) and Florence *et al.* (1995), as the rocks in question occur near AMCG plutons. Consequently, there is uncertainty as to whether a pre-1140 Ma event, at least in the central Highlands, is due to regional or contact metamorphism, or both. To add to this uncertainty, Alcock & Muller (1999) also suggested that granulite-facies metamorphism and deformation in the region were roughly coincident with anorthosite emplacement, which may have occurred in two or more pulses. In this scenario, the measured pressures of 7–8 kbar represent the level of anorthosite emplacement in the crust, not subsequent crustal thickening. Thus, high-grade metamorphism occurred discontinuously over the period 1140–1060 Ma. This could be one explanation for the wide range in ages of metamorphic garnet, monazite and zircon obtained from the Highlands (*e.g.,* Mezger *et al.* 1991). In contrast to deep emplacement, Spear & Markussen (1997) suggested that the anorthosites were emplaced at shallower depths (3–4 kbar), and were subjected to slow cooling between 1130 and 1100 Ma, after which time thickening of the crust occurred, resulting in the 1090–1060 Ma granulite-facies metamorphism.

Titanite ages from the Highlands range from 1014 to 1050 Ma in mafic rocks and from 982 to 989 Ma in marbles (Mezger *et al.* 1991, 1992), either indicating multiple generations of titanite or a lengthy period of cooling following peak conditions of metamorphism. ^{40}Ar–^{39}Ar hornblende plateau ages from the region, although limited in number, suggest cooling through 500°C at roughly 930–920 Ma (Heizler & Harrison 1986, Onstott & Peacock 1987)

Comparison with possible correlative rocks in Canada sheds little additional light on the history of the Highlands. The Morin terrane (part of the Central Granulite terrane) in Quebec may represent an extension of the Adirondack Highlands (*e.g.,* Davidson 1998a). The Morin terrane consists of the 1155 ± 3 Ma (Doig 1991) Morin anorthosite and surrounding charnockites, granulites and paragneisses. The bounding Morin and Labelle shear zones are in the granulite facies (730°C, 7.9 kbar: Hudon & Martignole 1996; 725°C, 8.5 kbar: Indares & Martignole 1990b); deformation occurred prior to ~1070 Ma (Martignole & Friedman 1992). Thus the Morin terrane and the Adirondack Highlands are similar in terms of the age of AMCG plutonism and the timing of regional granulite-facies metamorphism and deformation.

McLelland *et al.* (1996) noted that ages from AMCG suite plutonic rocks and granulite-facies metasediments in the St. Maurice area of Quebec, particularly the Shawinigan domain (Corrigan & van Breemen 1997) are similar to those in the Adirondack Highlands. A notable difference in the St. Maurice area is the geochemical and geochronological evidence for basement consisting of either a ~1450 Ma island arc that was subjected to ~1370 Ma regional metamorphism, or a ~1380 Ma continental arc (*e.g.,* Corrigan & van Breemen 1997); this history is analogous to that proposed for the Laurentian margin (*e.g.,* Culshaw *et al.* 1997). The similarity in the 1180–1040 Ma metamorphic and plutonic history between the Adirondack Highlands and the St. Maurice area, however, may indicate that the AMCG magmatism and subsequent high-grade metamorphism were widespread in the southern Grenville Province, and were superimposed on several earlier-formed and amalgamated belts.

TABLE 2. SUMMARY OF METAMORPHIC AND DEFORMATION HISTORY OF THE SOUTHERN PROVINCE PROPOSED BY VARIOUS AUTHORS

Events	Timing	Orogeny	Source
1) early deformation, low-grade metamorphism (<middle greenschist)	pre-Nipissing gabbro (2.2 Ga)	Penokean	Card (1964, 1978)
2) peak metamorphism, folding in higher-grade areas	pre-SIC (1.85 Ga)		
3) post-metamorphism deformation, cross-folding, alteration of porphyroblasts	1.85–1.7 Ga		
1) deformation and folding 2) metamorphism 3) late deformation, crenulation	post-SIC (1.85 Ga)	Penokean	Brocoum & Dalziel (1974)
North of Murray Fault			
low grade, upright folds	affects Nipissing and SIC (<1.85 Ga)	pre- and post-Penokean, no Penokean (1.87–1.82 Ga events)	Jackson (1997, 1998)
South of Murray Fault			
1) folding 2) *D*1 structures, pre- to syn-peak metamorphism	pre-Nipissing (>2.2 Ga)		
3) *D*2 structures, folds that deform isograds.	post-Cutler (<1.75 Ga)		
post-1.85 Ga metamorphism, early amphibolite facies in Elsie Mountain Fm. could be SIC contact effect or regional metamorphism	post-SIC (1.85 Ga)	Penokean	Fleet *et al.* (1987)
1) folding, faulting, high-grade metamorphism in metavolcanic rocks in the Sudbury area	2.47–2.33 Ga; using revised age of Creighton pluton, age range is reduced to ~2.47 Ga	Blezardian	Riller *et al.* (1999)
2) shear zones, fabric development, retrogression at greenschist facies	post-SIC (<1.85 Ga)	Penokean	
1) pre-Nipissing folding			Zolani *et al.* (1984)
2) intrusion of Nipissing gabbro	2.2 Ga		
3) compression and burial to 15–20 km, peak metamorphism	~1.9 Ga		
4) thrusting, folding, cleavage development			

SIC: Sudbury Igneous Complex.

Overview and Discussion

Southern Province

As summarized in Table 2, the timing of metamorphism in the Southern Province has been a topic of considerable debate. Card (1964, 1978) considered the deformation and metamorphism to be Penokean, post-intrusion of the Nipissing gabbro, but pre-intrusion of the Sudbury Igneous Complex. Brocoum & Dalziel (1974), however, argued that metamorphism took place during a Penokean orogenic event that postdated intrusion of the Sudbury Igneous Complex. The work of Fleet *et al.* (1987) indicates that regional metamorphism did indeed occur after emplacement of the Sudbury Igneous Complex, which is also consistent with the timing of metamorphic and structural events outlined by Jackson (1997) for the Southern Province west of Sudbury. Jackson (1997) proposed that the peak conditions of metamorphism occurred soon after emplacement of the Nipissing gabbro (*i.e.*, ~2220 Ma). Rb–Sr ages on Nipissing gabbro and Gowganda Formation argillites from the Gowganda area are 2116 ± 54 Ma and 2240 ± 174 Ma, respectively (Fairbairn *et al.* 1969). These ages are consistent with Jackson's (1997) interpretation that regional metamorphism closely followed the emplacement of the Nipissing gabbro.

Several outstanding problems remain with respect to the metamorphic history preserved in the Southern Province. A partial list follows: 1) Understanding the differences in structural style and metamorphic grade north and south of the Murray fault zone. 2) Although some folding occurred pre-emplacement of the Nipissing gabbro (*e.g.*, Card 1978), it is unclear if it this folding was associated with regional metamorphism, even at low grade. 3) It has been suggested that more than one age of emplacement of gabbro is present, that is, some intrusions may be pre-folding, and others post-folding (*e.g.*, Jackson 1997). 4) As previously noted, the age of metamorphism is poorly constrained, and in the Sault Ste. Marie – Sudbury area, an episode of regional metasomatism at ~1700 Ma makes the interpretation of isotopic ages problematic. Metamorphism is post-Nipissing gabbro emplacement (~2220 Ma), but older than the 1747 Ma Cutler batholith. The latter postdated D_2 structures that deform metamorphic isograds (Jackson 1997), thereby placing a lower limit on the age of metamorphism. 5) The possibility of multiple events must also be considered, which may be one reason for the wide range in interpretation of the deformational and metamorphic history of the Southern Province as summarized in Table 2. For example, were Huronian metavolcanic rocks and the Murray and Creighton plutons metamorphosed and deformed at ~2450 Ma, as suggested by Riller *et al.* (1999)? If so, did this occur prior to deposition of the main Huronian sedimentary sequence? Was the main Huronian sedimentary sequence metamorphosed between 2220 and 1850 Ma, subjected to deformation at ~1750 Ma, and metasomatism at ~1700 Ma? Or were regional metamorphism and metasomatism solely restricted to the interval 1750–1700 Ma? 6) Regardless of the age of metamorphism, what was the source of heat for metamorphism? Crustal extension over a mantle plume related to Nipissing magmatism was suggested by Jackson (1997), or was there some other mechanism? Resolution of these questions is beyond the scope of this paper. Nonetheless, it is because of these difficulties that on the metamorphic map (Fig. 2), Southern Province metamorphism is given a broad range in age (Geon 22–18), and that areas of possible overprinting or younger metamorphism are shown during Geon 17.

Grenville Province

Unlike the Superior (Easton 2000) and Southern provinces in Ontario, the metamorphic and tectonic history of the Grenville Province in Ontario is relatively well understood, as recently summarized by Carr *et al.* (2000). In this section, the metamorphic history is briefly summarized. Three main pulses of Grenvillian metamorphism are present in the Grenville orogen in Ontario and New York, at ~1245 Ma, 1175–1160 Ma, and 1070–1050 Ma.

The ~1245 Ma metamorphic event is apparently restricted to the Composite Arc Belt. It exhibits a pattern similar to that found in greenstone belts of the Superior Province (Easton 2000), namely, regional greenschist- to lower amphibolite facies, with the amphibolite facies locally developed around major intrusions. This event was syn- to post-assembly of the various arc elements that constitute the Composite Arc Belt, and was likely associated with voluminous gabbroic and granitic magmatism in the belt at 1250–1242 Ma (*e.g.*, Carr *et al.* 2000).

Metamorphism at 1175–1160 Ma occurred mainly under conditions of the upper amphibolite to granulite facies, and affected rocks of the Frontenac–Adirondack belt, the southern margin of the Composite Arc Belt, and rocks of the Parry Sound domain, as previously described. This event may be present in rocks of the northern margin of the Composite Arc Belt, within the Central Metasedimentary Belt boundary thrust zone (CMBbtz). This suggestion is based mainly on geochronological evidence (*e.g.*, Carr & Berman 1997, McEachern & van Breemen 1993), although Peck & Valley (2000) present oxygen isotope data from this area that suggest a polymetamorphic history. What is unique about this event is that in the Frontenac and Parry Sound domains, early metamorphic assemblages were minimally affected by subsequent thermal events, with ^{40}Ar–^{39}Ar hornblende ages of ~1120 Ma being preserved in Frontenac domain, and ages of ~1030–1000 Ma preserved in the Parry Sound domain (Wodicka *et al.* 2000). In contrast, in other parts of the orogen, including the CMBbtz and the Adirondack Highlands, the 1175–1160 Ma event was absent or largely obliterated by subsequent regional metamorphism, as previously described in this paper. The current data suggest that the 1175–1160 Ma event was largely restricted to the southern part of the orogen (parts of the Composite Arc Belt and the

Frontenac–Adirondack belt). Carr *et al.* (2000) related this metamorphic event to coalescence of these two tectonic elements at 1190–1180 Ma, possibly with some limited interaction with the southern margin of Laurentia and the Laurentian margin. A combination of crustal thickening and the generation of felsic and anorthosite-producing magmas are considered to be the heat sources responsible for this event (*e.g.,* Carr *et al.* 2000). Sudbuction rollback has been suggested as a heat source for this event in rocks of the Parry Sound domain (Wodicka *et al.* 2000).

Metamorphism between 1070 and 1050 Ma across the orogen in Ontario is associated with major thrusting and tectonic assembly of all three tectonic elements of the orogen: Laurentia and the Laurentian margin, the Composite Arc Belt and the Frontenac–Adirondack belt. The heat source for this metamorphic event is attributed to thickening of the crust. Carr *et al.* (2000) have suggested that the extreme ductility seen in rocks within Laurentia and the Laurentian margin might reflect some preheating of the crust to 400–500°C due to mantle-plume activity associated with the Midcontinent Rift. In higher-grade or more slowly exhumed parts of the orogen (*e.g.,* Mazinaw domain, Adirondack Highlands, southern parts of L3 domain), the 1070–1050 Ma metamorphism appears to be of prolonged duration, extending to as young as 1020 Ma, with a range of metamorphic ages reported (see descriptions in previous sections of this paper). As discussed in Mezger *et al.* (1993), Culshaw *et al.* (1997), Ketchum *et al.* (1998) and Carr *et al.* (2000), post-1040 Ma regional extension was an important factor in controlling the present distribution of exposed crustal levels within the orogen.

Within the Grenville Front tectonic zone, metamorphism does not occur before 1030 Ma, and persists to 990 Ma, at which time U–Pb isotopic systems closed and the rocks were rapidly exhumed (*e.g.,* Haggart *et al.* 1993, Krogh 1994, Murphy 1999, Corfu & Easton 2000). This episode of exhumation is documented along the entire length of the exposed Grenville Front, from Ontario to Labrador (Krogh 1994), indicating that it is an orogen-wide event separate from the 1070–1040 Ma metamorphism.

The pattern of prolonged heating noted above is also observed in the higher-grade parts of the Superior Province (Easton 2000), and may simply indicate that mid- to low-crustal levels subjected to conditions of the upper amphibolite to granulite facies over a period of time continually re-equilibrate as a result of solid-state diffusion, repeated influx of heat through emplacment of magma, and the passage of fluids through the lower crust. Moser *et al.* (1996) proposed a model of arc–continent collision, and crustal delamination and extension, to explain the observed timing and distribution of metamorphism within the southern Superior Province, but whether this model is applicable to some aspects of Grenville Province metamorphism remains to be demonstrated. The observed pattern of heating within the Grenville Province begs the question of defining "orogenic pulses" (*e.g.,* Rivers 1997), especially when linking data from lower- and higher-grade domains. Do the high levels of the orogen record a pulse-like tectonothermal history, with lower levels presenting a more continuous history? Resolution of this question awaits an improved three-dimensional understanding of the architecture of the orogen.

With respect to identifying problem areas within the western Grenville Province requiring future study, the following suggestions are offered: 1) The metamorphic history of much of the L2 domain is poorly known. Parts of this region contain extensive areas of Archean rocks, and it is possible that early metamorphic events may be preserved in parts of this region. 2) The extent of ~1450 Ma granulite-facies metamorphism in L2 needs to be further delineated. 3) The granulites described by Davidson *et al.* (1982) southwest of Port Loring need to be studied in greater detail. 4) Eclogitic rocks in the Mattawa area need to be examined. 5) The metamorphic and tectonic history of the eastern half of L3 domain is poorly known. This area appears to contain less migmatite and more metaplutonic rock than areas to the west; does metamorphic history differ as well? 6) The driving forces for regional metamorphism within the Grenville Province remain problematic. Crustal thickening alone cannot account for the high grades of Grenvillian metamorphism. If the orogen was at least 60 km thick, some of the high-grade rocks were at mid-crustal levels and >700°C at the time of metamorphism. If a mantle source of heat is invoked, then there should have been massive melting of the lower crust and plutonism. The heat source may have been in the orogenic crust itself (*e.g.,* radiogenic heating); there are no heat-production data on the Central Gneiss Belt (*e.g.,* the ~1450 Ma plutonic rocks), however. 7) Within the Composite Arc Belt, the overprinting of ~1170 Ma metamorphism on rocks of the Sharbot Lake domain needs to be better documented. 8) Similarly, within the Frontenac–Adirondack belt, overprinting of Highlands metamorphism on the Lowlands needs to be better documented. 9) Metamorphism of marbles in both the Composite Arc Belt and the Forntenac–Adirondack Belt needs to be better documented, given that these rocks underlie large areas of these belts, and that these rocks host a variety of metallic and nonmetallic mineral deposits, many of which owe their origin to contact and regional metamorphism.

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