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METAMORPHISM OF THE CANADIAN SHIELD, ONTARIO, CANADA. I. THE SUPERIOR PROVINCE

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Abstract

This paper is an outgrowth of the compilation of metamorphic information for the Canadian Shield in Ontario, and complements the recently compiled Metamorphic Map of the Canadian Shield. The paper contains a summary of the Archean metamorphic history of the Superior Province in Ontario on a subprovince basis, with an emphasis on the limitations of the existing data and the reasoning applied in extrapolating metamorphic boundaries. Little of the early metamorphic history of the Superior Province (pre-2715 Ma) is preserved, but there is local evidence for events at 2870–2850, 2810, and 2730 Ma within some of the older blocks of crust. The present distribution of metamorphic grade and age of metamorphism largely reflects pan-Superior events in the interval 2710–2640 Ma that occurred subsequent to coalescence of a system of island arcs, back-arcs, oceanic plateaus and microcontinents between 2720 and 2690 Ma. The distribution of metamorphic events and facies is the result of three interrelated patterns. 1) There is a relationship between subprovince type, metamorphic grade and age, with granite–greenstone subprovinces generally preserving older greenschist- to lower-amphibolite-facies events, metasedimentary subprovinces preserving younger middle-amphibolite- to granulite-facies events, and high-grade gneiss subprovinces preserving the youngest events. 2) Discrete metamorphic episodes between 2720 and 2640 Ma are associated with periods of major plutonism. 3) There is a pattern of increasing complexity of metamorphic history with increasing metamorphic grade. The present distribution of metamorphic facies in the Superior Province also was influenced by uplift, tilting, and erosion during the Paleoproterozoic. The timing of lode-gold and rare-element-pegmatite mineralization within the Superior Province corresponds closely with metamorphic evolution; it is consistent with models whereby gold-bearing fluids and pegmatite-forming melts develop, in part, as a result of granulite-facies metamorphism of the lower crust.

Keywords: metamorphism, Archean, Superior Province, granulite, amphibolite, greenschist, mineralization, Ontario.

Sommaire

Cet article résulte de la compilation d'information à propos du métamorphisme des roches du Bouclier Canadien en Ontario, et sert de complément à la nouvelle carte métamorphique du Bouclier Canadien. Cet article contient donc un résumé de l'évolution métamorphique de la province du Supérieur en Ontario et, tour à tour, de chaque sous-province; de plus, il souligne les lacunes dans les connaissances actuelles et le raisonnement qui a servi pour l'extrapolation des contacts entre sous-provinces. Très peu de détails ont survécu à propos de l'évolution métamorphique précoce de la province du Supérieur (avant 2715 Ma), mais il subsiste à l'échelle locale des signes d'événements à 2870–2850, 2810, et 2730 Ma au sein des socles plus anciens. La distribution actuelle des faciès métamorphiques et l'âge des épisodes de métamorphisme sont développés à l'échelle du Supérieur en entier sur l'intervalle 2710–2640 Ma, suite à la coalescence d'un système d'arcs insulaires, d'arrières arcs, de plateaux océaniques et de microcontinents entre 2720 et 2690 Ma. La distribution d'événements métamorphiques et des faciès résulterait de trois schémas interreliés. 1) Il y a une relation entre le type de sous-province, le degré de métamorphisme et l'âge, les ceintures montrant l'association granite – roches vertes conservant en général les secteurs plus anciens métamorphisés au faciès schistes verts et amphibolite inférieur, les sous-provinces à séquences métasédimentaires conservant l'évidence d'événements plus jeunes dans le faciès amphibolite moyen jusqu'au faciès granulite, et les sous-provinces à dominance de gneiss à degré de métamorphisme intense témoignant des événements les plus récents. 2) Les épisodes métamorphiques distincts entre 2720 et 2640 Ma sont associés à des périodes de plutonisme important. 3) Il y a progression en degré de complexité de l'évolution métamorphique à mesure que l'intensité du métamorphisme augmente. La distribution des faciès dans la province du Supérieur a aussi été influencée par des épisodes de soulèvement, d'inclinaison, et d'érosion au cours du Paléoprotérozoïque. La répartition dans le temps des événements de minéralisation, menant à la déposition de l'or en veines et aux venues de pegmatites granitiques enrichies en éléments rares dans le Supérieur, correspondrait étroitement à l'évolution métamorphique. En particulier, les fluides aurifères et les magmas felsiques propices à cristalliser sous forme de pegmatites se développeraient, au moins en partie, suite à un événement de métamorphisme dans le faciès granulite dans la croûte inférieure.

(Traduit par la Rédaction)

Mots-clés: métamorphisme, archéen, province du Supérieur, granulite, amphibolite, schistes verts, minéralisation, Ontario.

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Introduction

The only previous compilation of the metamorphic history of Ontario occurred over twenty years ago, resulting in the Ontario portion of the metamorphic map of the Canadian Shield (Fraser *et al.* 1978; referred to hereafter as the 1978 Map). During that project, data were compiled at 1:1 000 000 scale and published at 1:3 500 000 scale. Apart from five geographically localized contributions (Ayres 1978, Pirie & Mackasey 1978, Thurston & Breaks 1978, Jolly 1978, Card 1978) in *Metamorphism in the Canadian Shield* (Fraser & Heywood 1978), little record exists of the thought processes and the underlying database utilized in that compilation.

As outlined in Berman *et al.* (2000), a newly revised version of the Metamorphic Map of the Canadian Shield has been compiled (Berman, in prep.). This paper represents an outgrowth of the compilation of metamorphic information for the Canadian Shield in Ontario (Fig. 1). The purpose and organization of this paper are twofold. The first part briefly describes the Archean metamorphic

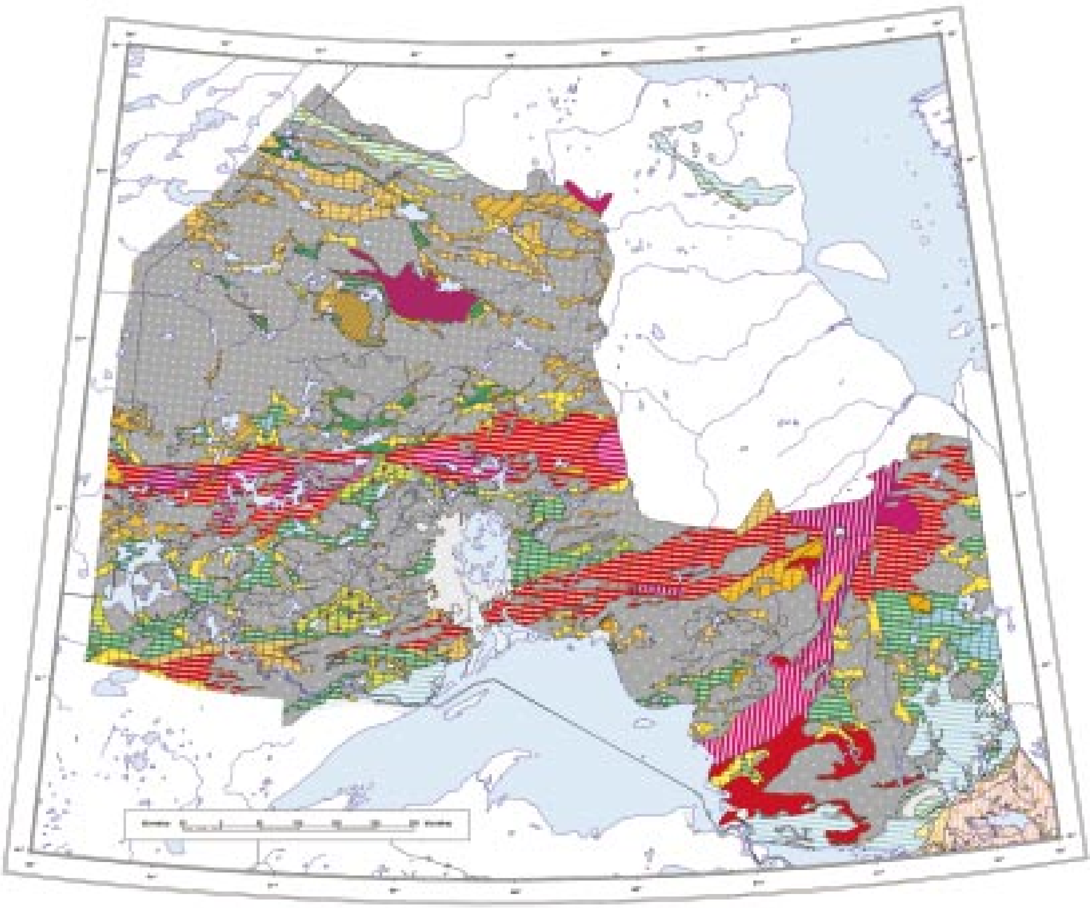
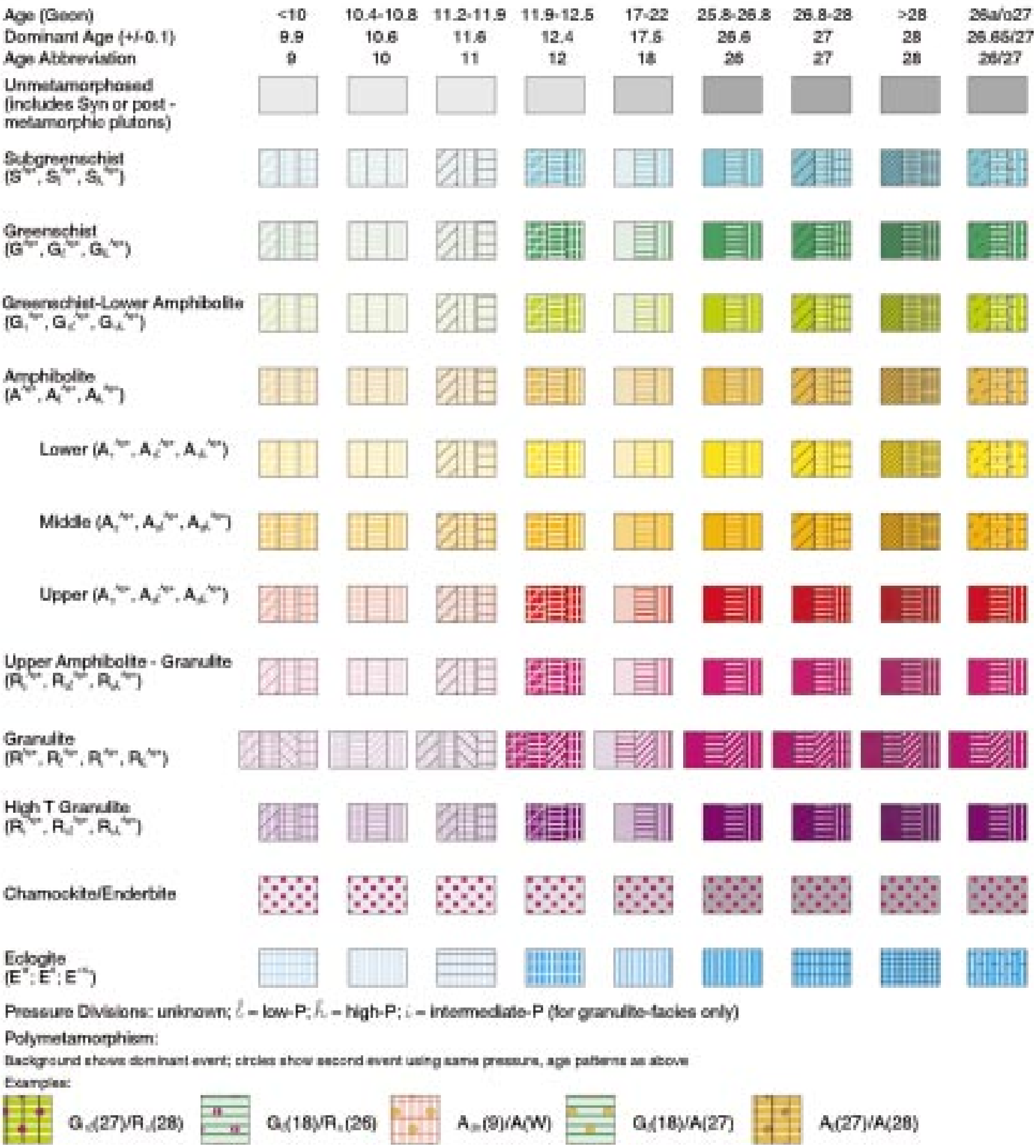


FIG. 1. Simplified version of the Ontario portion of the Tectonometamorphic Map of the Canadian Shield. A brief explanation of the legend follows; details can be found in Berman *et al.* (2000) and Berman (in prep.). "Age" refers to the main age divisions shown on this map; "age abbreviation" refers to the corresponding age units on the tectonometamorphic map of the Canadian Shield (Berman, in prep.), and used for description of polymetamorphic relationships in the legend. For each category of metamorphism, grade is portrayed in color. Lower grades are shown in cooler colors, higher grades, in warmer colors. Within each grade category, greenschist facies, for example, darker colors indicate older ages of metamorphism. Within each category of grade, except for the granulite facies, pressure is portrayed with thick lines as follows: low pressure: horizontal, medium pressure: vertical, unknown: diagonal or grid (if two colors are used age definition). In the case of the granulite facies, unsubdivided: no lines or hatched lines, low pressure: horizontal line, medium pressure: diagonal line, high pressure: vertical line. Overprinting relationships are shown with colored dots, with the dots representing the older metamorphic event, the background the younger event. For example, the first box on the left (number 863) represents a 990 Ma low-pressure, lower-amphibolite-facies event that overprints a 1000–1200 Ma amphibolite-facies event.

history of the Canadian Shield in Ontario, and explains how units are portrayed on the map, with an emphasis on the limitations of the existing data and the reasoning applied in extrapolating metamorphic boundaries between well-studied and poorly known areas within each subprovince. The second part discusses the constraints metamorphic history places on the tectonic evolution of the Superior Province in Ontario. The Proterozoic metamorphic history of the Canadian Shield in Ontario is presented in a companion paper in this issue (Easton 2000; hereafter referred to as Part 2). These two papers can be considered as a set of "marginal notes" for the Ontario portion of the Metamorphic Map of the Canadian Shield (Berman, in prep.).

Details of the compilation process of the Ontario portion of the Metamorphic Map are given in Easton (1999), with details of the legend design and terminology conventions described in a companion paper in this volume (Berman *et al.* 2000). The legend is akin to that used on the 1978 Map, in that it portrays metamorphic



units based on the metamorphic facies (subgreenschist, greenschist, amphibolite, granulite) and the presence or absence of overprinting relationships. Although a variety of metamorphic types have been identified in the better-studied parts of the Canadian Shield (*e.g.*, seafloor metamorphism, contact metamorphism, regional metamorphism, burial metamorphism, *etc.*), for the most part, the Metamorphic Map of Ontario (and the Canadian Shield) emphasizes the distribution of regional metamorphic events. Where recognized, other metamorphic types are noted in the descriptive sections of both papers.

A digital version of the 1:1 000 000 scale Tectonic Assemblages of Ontario map (OGS 1992a, b, c, d) was used as the base map. No attempt was made to update the base map to incorporate geological information acquired since publication, although such information was used in compilation. This decision was made to limit the scope of the project and to ensure that the metamorphic map was consistent with existing *Geology of Ontario* products.

Carmichael (1978) introduced the concept of metamorphic bathozones and bathograds as a measure of the depth of regional post-metamorphic uplift and erosion. This concept is used herein to complement the limited quantitative barometry currently existent for the Superior Province. Where quantitative information about P–T is presented herein, the equivalent bathozone is given in brackets for convenience; it should be recognized, however, that pressures derived from quantitative barometry and mineral assemblages are not directly equivalent. A bathograd is a pressure-sensitive metamorphic isograd based on an invariant mineral assemblage (Carmichael 1978). Using five bathograds, Carmichael (1978) defined six bathozones, numbered from 1 to 6 in order of increasing pressure, as illustrated in Figure 2.

Superior Province: Introduction

Space does not allow for a detailed overview of the geology of the Superior Province, which can be found in Thurston *et al.* (1991b), Williams *et al.* (1992), and Card & Poulsen (1998). In brief, the Meso- and Neoarchean Superior Province forms the cratonic core of North America, and is surrounded and truncated by Proterozoic orogenic belts (Hoffman 1989). The Superior Province has been traditionally subdivided into structural subprovinces on the basis of tectonic trends (Stockwell 1964), taking into account variations in lithology, metamorphism, and geophysical characteristics (*e.g.*, Card & Ciesielski 1986). Four distinctive types of subprovinces are recognized: granite–greenstone, metasedimentary, plutonic, and high-grade gneiss (*e.g.*, Card & Ciesielski 1986, Thurston *et al.* 1991b, Williams *et al.* 1992, Card & Poulsen 1998). Subprovince boundaries are commonly zones of structural and metamorphic contrast, and many are marked by east–west faults. Figure 3 illustrates the tectonic subdivisions of the Canadian Shield in Ontario utilized in this paper, slightly modified from those of Thurston *et al.* (1991b). Table 1 summarizes the constraints on the timing of peak metamorphism, P–T conditions and key mineral assemblages for each of the major subdivisions of the Superior Province described below.

Northwest Superior Province: Tectonic Framework

The tectonic framework of the Northwest Superior Province is currently in a state of flux (*e.g.*, Thurston 1999). The region (Fig. 3) had been divided into classical subprovinces, but, more recently, attempts have been made to subdivide it into several terranes (*e.g.*, Thurston *et al.* 1991a, 1998, Stott & Corfu 1991). Because of this lack of consensus on tectonic nomenclature, in this paper, the Northwest Superior Province is defined as the

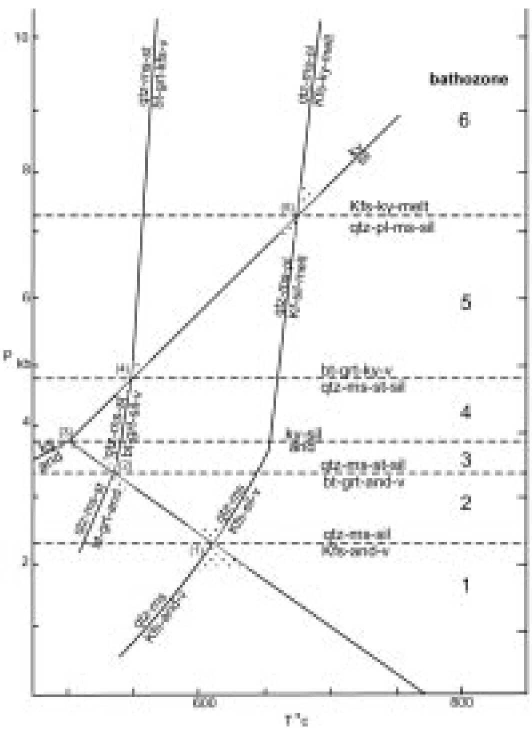


FIG 2. Idealized P–T phase diagram for part of the system SiO_2–Al_2O_3–FeO–MgO–Na_2O–K_2O–H_2O showing the five invariant points on which bathograds are based, and the resultant bathozones (after Carmichael 1978). Diagnostic phase-assemblages that constrain each bathograd are labeled on the appropriate sides of the hachured isobaric line. Reproduced with permission of the *American Journal of Science*.

area including the Uchi, Berens, Sachigo and Winisk subprovinces of Card & Ciesielski (1986) (Figs. 1, 3). This area is discussed as a single entity because of the general uniformity of metamorphism across it.

Within the better-studied Uchi subprovince, Stott & Corfu (1991) recognized four main tectonic elements: the North Caribou terrane, the Pickle terrane, the Woman assemblage autochthon, and the south Uchi parautochthon. The metamorphic history of the rest of the Northwest Superior Province is generally consistent with the tectonic history of the Uchi subprovince outlined by Stott & Corfu (1991), namely: 1) the North Caribou terrane contains remnants of a Geon 28 metamorphic event that has largely been overprinted by ~2705 Ma metamorphism, 2) the Pickle terrane shows evidence of ~2730 Ma metamorphism, and 3) the Woman autochthon and the south Uchi parautochthon were assembled at some time after 2713 Ma, with peak metamorphism occurring prior to intrusion of the Blackstone pluton at 2702 ± 1 Ma.

FIG. 3. Tectonic subdivision of Ontario used in this paper (after Thurston 1991).

TABLE 1. SUMMARY OF METAMORPHIC INFORMATION FOR THE SUPERIOR PROVINCE, ONTARIO

Subprovince or Region	Subarea or Event	Age of peak metamorphism	Bathozone	Temperature estimate	Pressure estimate	Key minerals[1]	Facies
central and eastern NW Superior		~2715 Ma	2	540°C	~3.4 kbar	Crd–And–St	gs-la
Berens River		~2715 Ma	3–4	630°C	5 kbar	St–Sil	la-ua
Wachusk metamorphic domain	western	2870 ± 2 Ma	3/43	540°C	~3.4 kbar	Crd–And–St, And–Sil, early Bt–Ky–St	gs-la
	eastern	2856 ± 2 Ma	unknown			unknown	ua-gr
Uchi	M1	~2730 Ma	2	550–600°C	>2.5, <3.4 kbar	Crd	sg-la
	M2	2705 ± 3 Ma	2	500–540°C	<3.4 kbar	Grt–And–St	gs-m
Winisk		Geon 27?	unknown				ua-gr
Kenyon		~1800 Ma	unknown				gs?
English River	dynamic	2693–2691 Ma	4	650–750°C	4.7–5.4 kbar	Opx	gs-gr
	static	2678–2669 Ma	4	650–750°C	4.7–5.4 kbar	Opx	gr
Bird River		unknown	unknown	500–550°C			la-ma
Winnipeg River		2710–2678 Ma	4	700–750°C	4.0–6.0 kbar	Opx	ma-gr
Wabigoon	eastern, central, western	2701 ± 3 Ma	2	500–600°C	<3.4 kbar	Grt–Crd–St	gs-ma
	central	2809 Ma		>550°C			a
Wawa	low grade	2689–2684 Ma	2?	500–600°C	<3.4 kbar		gs-ma
	M1-high grade	2688–2684 Ma	5	600–650°C	6–7 kbar	Ky	ma
	M2-high grade	~2678 Ma	3–4	510–600°C	4–5 kbar	Grt–St–Sil	gs-ua
Quetico	M1-western	2702–2688 Ma	5?			Ky–St–Bt	a
	M2-western	2671–2665 Ma	1	500°C	2.5 kbar	And	gs-ma
	M2-central	2671–2665 Ma	2	620°C	3.3 kbar	Grt–And	gs-ua
	M2-eastern	2671–2640 Ma	3–4, locally 5?	700–780°C	5.4–6.1 kbar	St–Sil	la-gr
Abitibi	M1-central	2700–2690 Ma	1–2	<550°C	<3.4 kbar	Prh–Pmp, Ep–Act	sg-ma
	M2-central	~2670–2660 Ma	1–2	<500°C	<3.4 kbar	Prh–Pmp, Ep–Act	sg-la
	Ramsey–Algoma complex	2674–2669 Ma	3–4	>550°C			gs-ua
	Levack	2647–2642 Ma	5	750–800°C	6–8 kbar	Opx–Grt–Bt	gr
Pontiac		~2670–2660 Ma	3–4	>550°C	5.4–6.1 kbar	St–Sil	gs-ua
Kapuskasing structural zone	Fraserdale–Moosenee block		6		8–10 kbar	Opx	ua-gr
	Groundhog River block	2657–2648 Ma	6		7–9 kbar	Opx	ua-gr
	Chapleau block	mainly 2663–2630 Ma	6	720–850°C	8–11 kbar	Opx	ua-gr
	Wawa gneiss domain	2685 in west to 2660 in east	5		5–6 kbar		ua-gr

[1] Minerals noted in this column do not necessarily reflect bathograd-defining assemblages.
Abbreviations: And: andalusite, Act: actinolite, Bt: biotite, Crd: cordierite, Ep: epidote, Grt: garnet, Ky: kyanite, Opx: orthopyroxene, Prh: prehnite, Pmp: pumpellyite, Sil: sillimanite, St: staurolite, a: amphibolite facies, gr: granulite facies, gs: greenschist facies, la: lower-amphibolite facies, ma: middle-amphibolite facies, sg: subgreenschist facies, ua: upper-amphibolite facies.

NORTHWEST SUPERIOR PROVINCE: EARLY NEOARCHEAN (GEON 28) METAMORPHISM

The core of the Sachigo subprovince, centered around North Caribou Lake, contains an area of Geon 28 metamorphic rocks that were apparently sheltered from Geon 27 metamorphism (Figs. 1, 3). This area is designated herein as the Wachusk metamorphic domain, and encompasses part of the North Caribou terrane of Thurston *et al.* (1991a).

The western half of the Wachusk metamorphic domain is better documented, and consists of gneiss and metaplutonic rocks, partly enveloping the North Caribou Lake, Horseshoe Lake and Upper Windigo Lake greenstone belts, each containing a collage of supracrustal rocks deposited in the interval 2980–2870 Ma. These greenstone belts are characterized by extensive areas of greenschist-facies rocks, with narrow (<1 km wide) zones of amphibolite-facies rocks at their margins. In contrast, the late Neoarchean (Geon 27) belts within the Northwest Superior Province are dominantly amphibolite-facies or greenschist-to-amphibolite transitional facies belts, with greenschist-facies rocks only being found in the cores of the largest belts. Older plutonic and supracrustal rocks also occur further to the west in the Favourable Lake and North Spirit Lake greenstone belts, and possibly in the Sandy Lake and Muskrat Dam belts (see summary in Tomlinson *et al.* 1999), but at present there is no evidence that these older supracrustal rocks were affected by Geon 28 metamorphism. As these westerly belts show the Geon 27, rather than Geon 28, metamorphic pattern, they are excluded from the Wachusk metamorphic domain.

In the Berens River area, there are metaplutonic rocks older than 2750 Ma, but it is unclear if all of this region was subjected to an older regional metamorphic event. Corfu & Stone (1998a) reported a belt of 2860 Ma gneissic tonalite located between the McInnes Lake and Hornby Lake greenstone belts; however, dark and pale types of titanite from a sample of these gneisses suggest isotopic resetting at ~2692 Ma during waning of Geon 27 metamorphism (Corfu & Stone 1998b). Corfu & Stone (1998a) noted that most of the 2750–2700 Ma plutons in the Berens River area contain zircon grains with cores, suggesting a wider distribution of older crust in the Berens River area. Older plutonic rocks, in the 2925–3000 Ma age range, have been previously described from the North Spirit and Favourable Lake areas (Corfu & Wood 1986, Corfu & Ayres 1991, Stevenson 1995), but evidence for an older, pre-Kenoran metamorphism has yet to be documented for this part of the Northwest Superior Province. On the map (Fig. 1), areas of older gneissic rocks that have likely been partly or fully overprinted by Kenoran metamorphism are identified, although at the moment, it cannot be determined if these gneisses were subjected to one, two, or more regional metamorphic events.

In the amphibolite-facies portion of the North Caribou Lake greenstone belt, Breaks *et al.* (1991) described a series of isograds, consisting of, from lowest to highest, St–Chl–Grt, And–Bt–Pl±Crd, and Sil+Crd. The breakdown of andalusite to form sillimanite suggests conditions locally higher than the 3/4 bathograd, that is, P–T conditions of ~3.5 to 3.8 kbar (T ~540°C). The rare occurrence of early-formed kyanite and staurolite in the belt suggests pressures >3.8 kbar at ~500°C, at least in the early stages of metamorphism. In the southern and western parts of the North Caribou Lake greenstone belt, the amphibolite-facies isograds are a contact-metamorphic effect of the North Caribou Lake batholith, which was emplaced at 2870 ± 2 Ma (deKemp 1987).

Evidence concerning the age of metamorphism in the eastern half of the Geon 28 domain (Fig. 1) is limited. The Schade Lake gneiss in the northeastern part of the North Caribou Lake belt has yielded a metamorphic age of 2856 ± 2 Ma (deKemp 1987). It is part of a larger gneiss domain that continues east to Wunnummin Lake and that is dominated by migmatitic rocks (Thurston *et al.* 1979). The gneiss domain is spatially associated with an aeromagnetic high and a paired Bouguer and vertical-gradient gravity anomaly, suggesting that much of it may have been subjected to upper-amphibolite- to granulite-facies metamorphism of unknown age. The presence of two Geon 11 carbonatite complexes (Big Beaver House and Schryburt Lake) within this block, but not elsewhere in the surrounding region, may also reflect the presence of an older, thicker block of crust underlying the gneiss domain.

NORTHWEST SUPERIOR PROVINCE: NEOARCHEAN (~2730 TO 2700 MA) METAMORPHISM

In the Uchi subprovince, most greenstone belts are characterized by low-pressure greenschist-facies mineral assemblages, with amphibolite-facies assemblages present mainly in areas close to major felsic plutons internal and external to the belts. Typical thermal aureoles are about 2 to 3 km wide (*e.g.*, northern and western Red Lake belt: Andrews *et al.* 1986; western Lake St. Joseph belt: Clifford 1969), although aureoles up to 10 km wide occur in some areas (*e.g.*, eastern Red Lake belt: Andrews *et al.* 1986; eastern Lake St. Joseph belt: Goodwin 1965; Miminiska – Fort Hope belt: Thurston & Breaks 1978). The thermal effect of external plutons is most apparent in the smaller greenstone belts, where only limited areas of greenschist-facies rocks are preserved (*e.g.*, Bee Lake belt: Stott & Corfu 1991). Within the greenstone belts, the common amphibolite-facies assemblages Grt–St–Bt–Chl and St–And–Grt–Bt±Chl suggest temperatures of 500–540°C and pressures <3.5 kbar (bathozone 2), consistent with P–T data from the Red Lake belt reported by Stone (1998). Metamorphism was mainly subsequent to the main episodes of regional, penetrative deformation across the Uchi subprovince that occurred between 2720 and 2710 Ma (Stott & Corfu 1991). The age of metamorphism is well constrained at 2705 ± 3 Ma, based on the ages of felsic plutons with well-developed thermal aureoles.

Peak metamorphism in the granite-dominated Berens River subprovince is difficult to constrain, in part because of the long-lived nature of the plutonism, which lasted from 2750 to 2705 Ma (Corfu & Stone 1998a), and which concluded with minor peraluminous granitic and sanukitoid plutonism in the period 2697–2686 Ma. Thus, local thermal anomalies were present in parts of this region at different times, reflecting the volume of magmatic activity at any particular time-interval. Corfu *et al*. (1998) suggested a minimum age for gneiss formation and metamorphism in the Favourable Lake greenstone belt of 2713 Ma, based on titanite data. A poorly constrained garnet age of 2716 ± 112 Ma from the McInnes Lake greenstone belt is consistent with this estimate (Corfu *et al.* 1998). Older, secondary titanite from the central Berens River subprovince yields a range of 2711–2706 Ma ages, possibly reflecting resetting associated with regional metamorphism. This age estimate lies between that of an older 2730 Ma metamorphism and the widespread 2700 Ma event in the Uchi subprovince. Because of the overlap of plutonic and metamorphic history, most plutons in the Berens River subprovince, and the northwest Superior Province as a whole, are shown as synmetamorphic on the metamorphic map (*e.g.,* Fig. 1).

Stone (2000) used the Al-in-hornblende geobarometer to show that there is a pattern of decreasing depth of emplacement with age for plutonic suites in the Berens River area, with older plutons emplaced at >15 km depth (4 kbar, ~730°C) and younger sanukitoid plutons being emplaced at depths of <5 km (1.9 kbar, ~770°C). In addition, Stone (2000) suggested that the greatest uplift occurred in the southern part of the area, including the Red Lake greenstone belt. This southern area contains the oldest primary "dark titanite" ages (2731–2699 Ma: Corfu & Stone 1998b). Secondary dark titanite shows the oldest ages (2711–2706 Ma) in the Red Lake area, with the youngest ages (2697–2686 Ma) adjacent to the Bear Head fault (Corfu & Stone 1998b); secondary pale titanite shows a similar distribution. The distribution of primary titanite ages appears to reflect regional cooling subsequent to the thermal peak during regional plutonism, whereas the distribution of ages based on secondary titanite and coeval apatite seems to reflect superimposed hydrothermal activity (Corfu & Stone 1998b).

Geon 27 metamorphism north and east of the Berens River subprovince and the Wachusk metamorphic domain is difficult to constrain because of the paucity of information. This region contains plutonic rocks similar to those recognized in the Berens River area, but they are undated. Greenstone belts, such as the Muskrat Dam belt, show a pattern of greenschist-facies cores and amphibolite-facies margins. St–And–Crd is the typical mineral assemblage in metasedimentary rocks (Ayres 1978), suggesting no higher than bathozone-2 conditions, consistent with limited P–T data that suggest conditions of 600°C and ~3.4 kbar from the western part of this area (Stone 1998). As the Muskrat Dam belt contains 2734 ± 2 Ma metavolcanic rocks (OGS 1992a), and 2715 ± 1 Ma granitic rocks (OGS 1992a), the timing of regional metamorphism may be similar to that observed in the Berens River subprovince, but at a higher level in the crust, based on the presence of andalusite–cordierite *versus* sillimanite–staurolite assemblages.

As noted by Ayres (1978), greenstone belts adjacent to and within the Berens River subprovince, including the North Spirit and Favourable Lake belts, contain dominantly middle-amphibolite-facies mineral assemblages, characterized by sillimanite–staurolite. These assemblages suggest bathozone-3–4 conditions in this part of the Northwest Superior Province, rather than the bathozone-2 conditions typical of the Uchi and North Caribou areas. This inference is consistent with thermobarometry on metasedimentary rocks from the Berens River and Favourable Lake areas that gives average P–T conditions of 630°C and 5 kbar (Stone 1998). Stone (1998) has also reported orthopyroxene from supracrustal rocks in the McInnes Lake and Favourable Lake belts, suggesting local pyroxene-hornfels- or granulite-facies conditions.

Although not distinguished on the map (Fig. 1), within the Pickle terrane (Uchi subprovince), there is geochronological and minor field evidence for two Geon 27 metamorphic events. The strongest evidence comes from the Pickle Lake greenstone belt. The 2740 ± 2 Ma Pickle Lake stock contains two fractions of titanite, both of which are dated at 2734 ± 4 Ma (Corfu & Stott 1993b), and the 2836 ± 3 Ma Quarrier gneiss contains 2716 ± 2 Ma titanite (Corfu & Stott 1993a). In addition, the Ochig Lake pluton has been dated at 2741±2 Ma (Corfu & Stott 1993b). Although titanite ages from metaplutonic rocks are difficult to interpret, since they could simply represent cooling ages, the fact that a similar age is recorded in a variety of metaplutonic units of varied age of emplacement suggests that the titanite ages may reflect a regional metamorphic event. Also compelling is that all these units either intrude or adjoin the Geon 28 Northern Pickle and Pickle Crow supracrustal assemblages (Stott & Corfu 1991), which have been subjected to low-pressure, amphibolite-facies metamorphism; in fact, the greenschist–amphibolite boundary in this belt essentially follows the assemblage boundary. Patterson & Watkinson (1984) estimated metamorphic conditions within the Northern Pickle assemblage at 600°C (Grt–Bt) and 5.5 kbar (Grt–Pl) (bathozone 4), higher than the <550°C and <4 kbar (bathozone 2) conditions estimated for amphibolite-facies rocks within the Woman autochthon and the south Uchi parautochthon. Further, the North Pickle assemblage at Pickle Lake was the locus of extensive hydrothermal alteration, folding, with zones of ductile deformation, and important gold mineralization (Stott & Corfu 1991). The titanite ages, the higher P–T conditions, the abundant zones of deformation, and the relationship of metamorphic grade to lithotectonic boundaries all suggest that the amphibolite-facies metamorphism present in the North Pickle assemblage could be ~2735 Ma in age.

A ~2730 Ma metamorphic event may be present within the North Caribou terrane of Stott & Corfu (1991). The 2750 ± 3 Ma Dobie Lake pluton, located north of the Meen–Dempster greenstone belt, contains 2734 ± 3 Ma titanite (Corfu & Stott 1993a) and has imposed a narrow contact-metamorphic aureole on adjacent greenstones, yet the pluton is weakly metamorphosed and contains a strain fabric (Stott & Corfu 1991). In the Red Lake greenstone belt, Menard *et al.* (1999) suggested the presence of an early greenschist- to lower-amphibolite-facies metamorphism at ~2731 Ma, the age of emplacement of the Little Vermilion Lake batholith. This suggestion is consistent with preservation in the northern Red Lake greenstone belt of ~2720 Ma hornblende ^{40}Ar–^{39}Ar ages (Layer *et al.* 1987) in rocks of the Hammel Lake batholith. The hornblende ages are similar to U–Pb zircon and titanite ages from the same sample, and show no obvious excess Ar signature, suggesting that in parts of this area, the metamorphic conditions at ~2705 Ma were lower than 500°C.

Within the central part of the Birch–Uchi greenstone belt, Thurston (1985) described a prograde sequence from subgreenschist- to middle-amphibolite facies, and tentatively identified the assemblage Cal–Ep–Tr–Act–Ab–Pmp within the lowest-grade part of the belt. The absence of prehnite in the assemblage may reflect pressures greater than 2.5–3 kbar. Most of the rocks within the subgreenschist zone contain the assemblage Ep–Ab–Chl–Cal (Thurston 1985), which could be indicative of lower-greenschist-facies conditions; but higher CO_2 activity would stabilize carbonate minerals at the expense of Ca–Al silicates (Liou *et al.* 1987), thereby explaining why prehnite or pumpellyite (or both) are not abundant in the lowest-grade part of the belt. The Birch–Uchi belt also is unusual in that the metasedimentary rocks locally contain Crd–Ms±Kfs±Grt±And (Thurston 1985), in contrast to the And–St assemblages typical of the southern Uchi subprovince (Stott & Corfu 1991). This difference suggests temperatures of ~550°–600°C, and pressures of <3 kbar for the central Birch–Uchi belt (bathozone 2). Thurston (1985) noted that the contact-metamorphic aureole of the Okanse Lake pluton overprints regional metamorphic minerals, such as cordierite, thus constraining the age of metamorphism between 2735 Ma, the age of youngest volcanism in the area, and 2701^{+2}_{-1} Ma, the age of the Okanse Lake pluton (Beakhouse *et al.* 1999).

Middle- to upper-amphibolite-facies conditions occur close to the English River subprovince boundary, and the regional map-pattern suggests that these higher-grade conditions could be the result of overprinting English River metamorphism (~2685 Ma) on the Uchi subprovince. The presence of an overprinting relationship remains to be fully documented.

Northwest Superior Province: Winisk Subprovince

Thurston *et al.* (1991a) defined the Winisk subprovince as consisting of all Archean rocks east of the Winisk River fault. The Winisk suprovince is underlain by clinopyroxene-bearing quartz monzonite and granodioritic gneisses, and is characterized by a regional aeromagnetic high with a curvilinear pattern, and a paired Bouguer and gravity-gradient high. Bostock (1971) interpreted the aeromagnetic pattern as representative of granulite facies, which is also consistent with the gravity signature. Various investigators (*e.g.*, Ayres *et al.* 1969, Thurston & MacFadyen 1992), primarily on the basis of aeromagnetic data, have correlated the Winisk subprovince with the Pikwitonei subprovince of Manitoba. This is the approach followed herein, but because of the limited amount of data from the Winisk subprovince, it is coded simply as Neoarchean upper amphibolite to granulite facies on the metamorphic map (Fig. 1). If equivalent to the Pikwitonei subprovince, metamorphic ages of 2700–2690, 2688–2676, and 2648–2637 (*e.g.*, Mezger *et al.* 1989) would be expected.

Northwest Superior Province: Kenyon Structural Zone

The Kenyon structural zone (Ayres *et al.* 1971) is characterized by a west–northwest-trending, moderate to strong, linear aeromagnetic pattern (Gupta 1991) that varies from 20 to 40 km in width, and stretches some 550 km from northeast of Knee Lake in Manitoba to the Winisk River fault in Ontario. Card & Sanford (1994) showed the zone as traversing Ontario but stopping at the Ontario–Manitoba border; however, the magnetic pattern characterizing the zone continues as far west as Knee Lake. As noted by Easton (1985), although this zone is likely underlain by Archean rocks, K–Ar biotite ages from the zone range from 1675 to 2300 Ma, some 200 to 825 million years younger than typical biotite ages from the Superior Province. In Ontario, the zone is bounded by two faults, the South Kenyon and North Kenyon faults. Stone *et al.* (1999) reported the presence of mylonites, with chlorite–epidote alteration, along the North Kenyon fault, but ductile deformational features along the South Kenyon fault; the latter may have served to localize emplacement of the Carb Lake carbonatite complex, dated at 1826 ± 97 Ma using K–Ar biotite (Sage 1987). Stone *et al.* (1999) reported the presence of hornblende tonalite and granite within the Kenyon structural zone near the Ontario–Manitoba border. Because of the resetting of K–Ar ages within the zone, it is shown on the metamorphic map (Fig. 1) as a zone of possible subgreenschist- to greenschist-facies Paleoproterozoic metamorphism that overprints the typical Geon 27 metamorphism of the Northwest Superior Province. It is possible that this overprinting occurred during Geon 18, and reflects distal effects of the Trans-Hudson orogen, which lies only 80 km to the north. Additional studies are necessary to determine the specific metamorphic history of the Kenyon structural zone, and the cause of the magnetic anomaly associated with it.

Northwest Superior Province: Sutton Inlier

Usage herein differs slightly from that in Thurston *et al.* (1991a), in that some exposed Archean rocks present beneath the Sutton Inlier are excluded from the Winisk subprovince. These Archean rocks lack the characteristic magnetic signature of the main mass of the Winisk subprovince, and Bostock (1971) described the rocks as undeformed granodiorite to monzogranite that intruded migmatitic granodioritic gneiss; both being affected by widespread chlorite alteration. On the metamorphic map (Fig. 1), these rocks are shown as Geon 27 amphibolite-facies rocks (based on the presence of the granodioritic gneiss), possibly overprinted by a Geon 18 subgreenschist-facies metamorphism or alteration related to the overlying Paleoproterozoic Sutton Group (see Part 2).

English River Subprovince

The history of metamorphic studies in the English River subprovince has been reviewed by Breaks (1991), from which this summary is adapted. Breaks (1991) suggested that the gross distribution of metamorphic zones within the subprovince was asymmetrical with respect to the English River – Winnipeg River subprovince boundary, with grade generally increasing from north to south toward this boundary. Notable exceptions to this pattern occur where lower-grade rocks of the Wabigoon and Bird River subprovinces abut the southern boundary of the English River subprovince, producing a more symmetrical metamorphic pattern, and where the middle-amphibolite-facies Melchett Lake greenstone belt lies isolated within upper-amphibolite metasedimentary migmatite and peraluminous granites. The regional distribution of metamorphic facies is commonly interrupted or telescoped along the subprovince boundary fault-zones, especially along the northern boundary (*i.e.*, the Sydney Lake – Lake St. Joseph fault). For example, in the Bee Lake – Rice Lake area, the Sydney Lake – Lake St. Joseph fault juxtaposes middle-amphibolite andalusite-bearing pelites against migmatite, a discontinuity corresponding to an abrupt rise in temperature of over 100°C and an increase in pressure of about 2 kbar (Campion *et al.* 1986). Discontinuities of similar magnitude occur in the Pashkokogan and Pakwash Lake areas. Relatively intact prograde metamorphic facies-series are preserved only in a few areas. Six metamorphic zones have been delineated between Lake St. Joseph and eastern Lac Seul by Breaks *et al.* (1978): Chl–Bt, St–Chl–Bt, Sil–Ms, Sil–Kfs, Crd–Grt–Kfs, and Opx.

Low-pressure granulite-facies metamorphism in the English River subprovince, first described by Breaks *et al.* (1978), occurs as ovoid nodes up to 100 km long by 25 km across, intermittently dispersed over a strike length of 500 km. The nodes occur near the English River – Winnipeg River subprovince boundary, and locally straddle it. Breaks (1991) interpreted these nodes as thermal culminations. The common granulite-facies assemblage is Crd–Grt–Qtz–Pl–Kfs. Crd and Opx coexist in some metasedimentary layers; more commonly, Opx or Di (or both) occur in layers of mafic and intermediate composition (Breaks & Bond 1993).

Breaks (1988) recognized two granulite-facies types in the eastern Lac Seul zone, namely, a dynamic (synkinematic) type and a static (late kinematic to postkinematic type). The dynamic type is the most areally extensive, and is characterized by orthopyroxene developed in association with D_2 fabrics. The static type is less extensive, and is characterized by a series of mineral assemblages containing orthopyroxene, biotite, plagioclase and iron oxides. Breaks (1988, 1991) suggested that five distinct metamorphic pulses occurred during the late kinematic to postkinematic history of the region, with the static-granulite type outlasting the emplacement of tonalite–granodiorite plutons. The timing of these pulses is discussed below.

P–T conditions during metamorphism have been estimated from mineral reactions (Breaks *et al.* 1978, Thurston & Breaks 1978) and by thermobarometric determinations (Henke & Perkins 1983, Henke 1984, Chipera 1985, Percival *et al.* 1985, Campion *et al.* 1986, Chipera & Perkins 1988) (Table 1). Pressure conditions of 4.7–5.4 kbar are reported for the eastern Lac Seul area (Baumann *et al.* 1984, Percival *et al.* 1985, Chipera & Perkins 1988), 4.0–5.0 kbar for the Whitewater–Mojikit lakes granulite zone (Percival *et al.* 1985), and 4.8–5.4 kbar for the Umfreville – Conifer Lake granulite zone (Pan *et al.* 1996). These values correspond to bathozone 4, and apparently show little regional variation in pressure conditions. Temperatures of 675–750°C have been reported from the eastern Lac Seul area (Baumann *et al.* 1984, Percival *et al.* 1985, Chipera & Perkins 1988), 680–730°C from the Whitewater–Mojikit lakes granulite zone (Percival *et al.* 1985), and 650–750°C for the Umfreville – Conifer Lake granulite zone (Pan *et al.* 1996). Pan *et al.* (1999) also obtained similar P–T values for rocks within the upper-amphibolite zone in the Umfreville – Conifer Lake area. The highest-grade rocks contain hercynite enveloped by cordierite, suggesting temperatures of 750°C (Henke 1984).

In the lower-grade marginal zones, amphibolite-facies conditions fall in the range of 550–650°C and

3.0–4.5 kbar (Breaks *et al*. 1978), consistent with thermobarometric determinations of 500–625°C and 4 kbar (Chipera 1985, Campion *et al.* 1986). Conditions of amphibolite-facies retrogression within the Umfreville – Conifer Lake granulite zone are similar at 550–600°C and 3–4 kbar (Pan *et al.* 1996).

Granulite-facies metamorphism (all types) in the eastern Lac Seul area is bracketed between 2692 Ma (age of pre-D_1 orthopyroxene-bearing leucosome in metasedimentary migmatite) and 2660 ± 20 Ma, the age of emplacement of late granite–granodiorite suite rocks during the waning stages of static granulite-facies metamorphism (Krogh *et al.* 1976). Dynamic granulite-facies metamorphism likely occurred at roughly 2693–2691 Ma, the age of: 1) metamorphic zircon in amphibolite at Miniss Lake (Corfu *et al.* 1995), 2) magmatic zircon in granitic leucosome in migmatite at Ramlin Lake and at eastern Lac Seul (Corfu *et al.* 1995), and 3) quartz syenite dikes with orthopyroxene selvages in the Umfreville – Conifer Lake granulite zone (Pan *et al.* 1999). Corfu *et al.* (1995) also suggested the presence of a 2678 Ma metamorphic event in the Lac Seul area, as well as a ~2669 Ma metamorphism, the latter event attributed to the late stages of static metamorphism (Breaks 1988, 1991). Tonalite-suite plutons emplaced between 2690 and 2700 Ma (*e.g.*, 2698 ± 2 Ma Fletcher Lake batholith, Corfu *et al.* 1995) are the only regionally distributed granitic rocks that could have supplied sufficient heat to effect high-grade metamorphism (Breaks 1991), yet these rocks are regionally metamorphosed, locally show evidence of partial melting, and are cut by the regional orthopyroxene isograd. Peraluminous granites within the subprovince are considered by Breaks (1991) to be a consequence of the dynamic granulite-facies metamorphism in the English River subprovince, rather than the cause. If Breaks (1991) is correct, then the exposed plutons in the English River subprovince are not the heat source responsible for regional metamorphism.

Winnipeg River Subprovince

The metamorphic history of the Winnipeg River subprovince is directly linked to that of the adjacent English River subprovince, the main difference being the largely metasedimentary character of the English River subprovince, in contrast to the dominant granitic character of the Winnipeg River subprovince.

The Winnipeg River subprovince is characterized by middle-amphibolite- to granulite-facies metamorphism, which affects the supracrustal rocks, rocks of the gneissic suite, and some of the plutonic rocks of the tonalitic suite (Beakhouse 1991). Although evidence for high-grade regional metamorphism is widespread within the subprovince, the peak conditions of metamorphism and regional variation in metamorphic grade are poorly understood owing to the paucity of mineral assemblages that constrain P–T conditions (Beakhouse 1991). The presence of *in situ* partial melts in mafic enclaves suggests temperatures in excess of 630°C (Wyllie 1977), and the presence of Cpx–Hbl amphibolites and Opx–Cpx–Hbl mafic granulites suggests that temperatures above 800°C were attained locally (Beakhouse 1991). At Kenora, Gower (1978) suggested peak conditions of metamorphism of 650–750°C and 4–7 kbar, whereas in the eastern Lac Seul area, P–T estimates of 700–750°C and 4–6 kbar were obtained by Perkins & Chipera (1985) and Chipera & Perkins (1988). Although many of the critical observations in the eastern Lac Seul area pertain to metasedimentary gneisses in the English River subprovince, granulite-facies assemblages straddle the subprovince boundary (*e.g.*, Breaks 1988). Furthermore, granulite-facies metamorphism in the Cedar Lake area developed under conditions comparable to those interpreted for the eastern Lac Seul area (Westerman 1977), with Ciceri *et al*. (1996) reporting temperatures of 654°C and 776–803°C from amphibolite- and granulite-facies rocks, respectively. In the Mystery Lake dome, Menard *et al.* (1997) reported peak conditions of metamorphism of 700°C and 4 kbar. Significantly higher pressures are recorded adjacent to the Winnipeg River – Wabigoon subprovince boundary near Dryden, namely, 600–650°C and 8–10 kbar at ~2685 Ma (Menard *et al.* 1997). Menard *et al.* attributed the P–T variations to peak conditions "freezing in" at different stages of syndeformational uplift along the boundary.

Tonalite gneisses from the eastern Lac Seul area formed between 3040 and 3100 Ma (Krogh *et al.* 1976, Corfu *et al.* 1995) and were metamorphosed at 2678 ± 2 Ma (Chamberlain Narrows batholith; Corfu *et al.* 1995), coincident with the age of metamorphism in the Cedar Lake area (2678^{+6}_{-4} Ma: Corfu 1988) and close to the age of static granulite-facies metamorphism in the adjacent English River subprovince. Zircon fractions from the Cedar Lake gneiss yield an age of 3170^{+20}_{-5} Ma (Corfu 1988). The Kenora gneissic suite in the Winnipeg River subprovince yields a U–Pb zircon age of 2840^{+20}_{-5} Ma (Corfu 1988). Similar ages are recorded from the Sioux Lookout gneiss complex (2889 ± 3 Ma: Corfu 1996) and from granodioritic rocks in the Cedar Lake area (2884 Ma: Cruden *et al.* 1997).

Ciceri *et al.* (1996) suggested that in the Kenora area, the 2710^{+5}_{-2} Ma Daniels Lake tonalite (Corfu 1988), and possibly other plutons of similar age in the Wabigoon subprovince, were responsible for regional metamorphism along the Winnipeg River – Wabigoon subprovince boundary. Card & Poulsen (1998) suggested the presence of three metamorphic events in the Winnipeg River subprovince, at ~2790, ~2705 and ~2680 Ma. In contrast, Beakhouse (1991) suggested that the difference in metamorphic ages between the northern and southern Winnipeg River subprovince may simply reflect differences in cooling related to greater depths at the time of regional metamorphism in the north, rather than two, or more, discrete episodes of metamorphism.

Bird River Subprovince

The Separation Lake greenstone belt at the northern margin of the Winnipeg River subprovince has been correlated with the Bird River greenstone belt in Manitoba, which was given subprovince status by Card & Ciesielski (1986). In Ontario, the subprovince is dominated by amphibolite-facies rocks, and the relationship between metamorphism in the Separation Lake greenstone belt and that in the adjacent Winnipeg River and English River subprovinces has yet to be established. Pan *et al.* (1999) noted that there is a significant discontinuity in metamorphic grade between the lower-amphibolite facies (500–550°C) in the greenstone belt and upper-amphibolite facies (650–750°C) in the southern English River subprovince. The Separation Lake greenstone belt also is the locus for one of the highest concentrations of rare-metal granitic pegmatite mineralization in the Superior Province (*e.g.*, Breaks & Tindle 1994, 1997, Tindle *et al.* 1998). Mineralization is genetically associated with the Separation Rapids pluton, which has been dated using monazite at 2646 ± 2 Ma (Larbi *et al.* 1999). The parent to the Separation Lake granite and related rare-metal granitic pegmatites are considered to have had a granulite-grade metasedimentary source (Breaks & Pan 1995, Pan & Breaks 1997, Pan *et al.* 1999). Consequently, coming to an understanding of the tectonic and metamorphic history of the Bird River and adjacent Winnipeg River and English River subprovinces may be important in terms of mineral exploration along this boundary and elsewhere in the Superior Province.

Wabigoon Subprovince

As with the northwest Superior Province, tectonic subdivision of the Wabigoon subprovince is in a state of flux, particularly with respect to the location of tectonic boundaries (*e.g.,* Tomlinson *et al.* 1999). For the purpose of this paper, the Wabigoon subprovince is divided into a Geon 27 domain located in the east and the west, dominated by supracrustal rocks, and a metamorphically more complex Geon 30–27 domain in the center (Figs. 1, 3).

In the western Wabigoon, pelitic rocks, although uncommon, contain low-grade assemblages of metamorphic minerals, and in the volcanic rocks, middle-amphibolite assemblages are the highest recorded (Pirie & Mackasey 1978). Variations in metamorphic grade are spatially associated with felsic plutonic rocks, with the highest metamorphic grades recorded adjacent to these intrusions. Metamorphic grade also increases at the margins of the Wabigoon subprovince, adjacent to the Quetico, Winnipeg River and English River subprovinces. Peak metamorphism appears to have post-dated major folding and deformation, as isograds cross major folds (Bartlett 1978, Pirie & Mackasey 1978, Poulsen *et al.* 1980, Poulsen 1984). Grt–Crd–St assemblages dominate amphibolite-facies domains in the western Wabigoon, suggesting bathozone-2 conditions (500–600°C, <3.4 kbar). This inference is consistent with mineral assemblages in hydrothermally altered volcanic rocks in the Sturgeon Lake belt of the western Wabigoon subprovince. These contain Ky–And, Sil–St (Lefebvre 1982), and Ky–And–Chd (Franklin *et al.* 1975) assemblages, suggesting conditions of <550°C and <4.0 kbar.

The age of metamorphism is tightly constrained at about 2701 ± 3 Ma, the age of the youngest grains of detrital zircon in western Wabigoon metasedimentary rocks (Davis 1995), the age of the oldest post-tectonic plutons (*e.g.,* Heronry Lake stock, Davis & Edwards 1986) and the age of titanite grains in pretectonic plutons (*e.g.,* Davis & Smith 1991).

Migmatitic metasedimentary rocks are absent from most of the western Wabigoon subprovince, except in the Dryden area, where they occur marginally disposed with respect to the Winnipeg River subprovince. Here, metasedimentary rocks display a northward progression from lower-greenschist to upper-amphibolite facies away from the Wabigoon fault; isograds generally parallel the fault (Bartlett 1978, Breaks *et al.* 1978, Breaks 1989). The presence of assemblages containing Crd, And and Sil (Bartlett 1978) indicates bathozone-2 conditions (500–600°C, <3.4 kbar); Menard *et al.* (1997), however, suggested that these isograds formed during a retrograde event at ~2647–2642 Ma, and that earlier (~2685 Ma) peak conditions were 650–700°C and 4–6 kbar, with the assemblage Grt–Crd–Bt–Pl and Sil–Kfs (bathozone 5). In contrast, along the southern margin adjacent to the Quetico subprovince, the occurrence of St–Grt–Sil and And (Poulsen 1984) indicates bathozone-3 conditions (500–600°C, >3.4 kbar).

An older segment of the crust, centered north of Lake Nipigon in the central Wabigoon, contains remnants of Geon 30, and possibly Geon 34–32, granitic rocks and supracrustal rocks (*e.g.*, Davis & Jackson 1988, Tomlinson *et al.* 1999). Further south, in the Lumby Lake area, there is evidence for a Mesoarchean metamorphic event. Here, the 3003 ± 5 Ma Marmion tonalite gneiss contains titanite, which dates an early metamorphic event at 2809 ± 1 Ma, whereas adjacent tuffs contain 2690 ± 5 Ma rutile, likely related to cooling after Geon 27 metamorphism. Younger volcanic sequences at high structural levels within the central Wabigoon subprovince appear to have been subjected to similar P–T conditions and timing of metamorphism as the eastern and western Wabigoon segments, *i.e.*, bathozone-2 conditions at ~2701 Ma. Preliminary data from the Sturgeon Lake – Obanga Lake area, however, suggest that at deeper structural levels in the central Wabigoon, metamorphism may have locally reached the granulite facies and is slightly younger (~2690 Ma Zrc, 2680–2670 Ma Ttn: Percival *et al.* 1999).

Quetico Subprovince

In this article, metasedimentary rocks east of the Kapuskasing structural zone are considered part of the Quetico subprovince, following the usage of Williams (1991) (Fig. 3). The intensity of metamorphism varies within the subprovince, such that rocks marginal to the subprovince tend to be at lower grade than in the interior. The lowest metamorphic grade is found along the northern boundary with the Wabigoon subprovince (Pirie & Mackasey 1978). Locally, subgreenschist- to greenschist-facies rocks occur along the southern boundary (Borradaile 1982), but typically, there is a rapid rise in metamorphic grade north of the Wawa subprovince, especially north of Manitouwadge, where a belt of metasedimentary granulites occurs within the Quetico subprovince close to, and parallel with, the northern margin of the Wawa subprovince (Coates 1968, Williams & Breaks 1989, 1990, Pan *et al.* 1994). As a result, grade distribution is asymmetrical, with the maximum in temperature and pressure occurring south of the central Quetico, locally coincident with the southern margin.

P–T conditions increase from west to east, for example, 500°C and 2.5 kbar at the Minnesota border west of Thunder Bay (Percival 1989), to 700–780°C and 5.4–6.1 kbar adjacent the Kapuskasing structural zone (Percival & McGrath 1986, Percival 1989). Typical conditions in the central region are on the order of 620°C and 3.3 kbar (Percival 1989). Granulites north of Manitouwadge yield 680–770°C and 4.4–6.4 kbar (Pan *et al.* 1994, Percival 1989). The regional variation in P–T can be ascribed to a relatively shallow level of erosion in the west (<10 km) and a deeper level in the east (>12 km) (Percival *et al.* 1985). Rocks located east of the Kapuskasing structural zone are believed to be generally at upper-amphibolite-facies conditions (Williams 1991).

Evidence for an earlier, medium-pressure, low-temperature, pretectonic or early syntectonic metamorphism comes from four areas within the subprovince. In the Atikokan region, and in northern Minnesota, both at the northern margin of the Quetico subprovince, an early M_1 metamorphic peak between D_1 and D_2 produced Ky–St–Bt assemblages (Ayres 1978, Tabor *et al.* 1989). Kyanite inclusions in plagioclase within Grt–Sil–Bt–Pl–Qtz schist near Raith, north of Thunder Bay, have been reported by Percival *et al.* (1985). Kehlenbeck (1976) also presented textural evidence for a polymetamorphic history along the northern margin of the Quetico subprovince north of Thunder Bay. Again, along the northern margin of the subprovince, in the Beardmore–Geraldton area (Williams 1989), amphibolite-facies conditions were attained prior to D_2 deformation, with the distribution of facies being structurally controlled within thrust-bounded panels (Williams 1991).

In contrast, the main phase of regional metamorphism (M_2), which produced the observed map- pattern (Fig. 1), occurred late syntectonically (Sawyer 1983, Williams 1991). The general sequence of isograds, based on the appearance of diagnostic assemblages in pelites, is Chl–Ms–Bt, Grt±And±Sil, Grt+Crd+Sil, *in situ* granitic leucosome, and Opx (Pirie & Mackasey 1978, Percival & Stern 1984). The common occurrence of Grt–And in metapelites in the western Quetico subprovince is diagnostic of bathozone 2 (<3.4 kbar; Carmichael 1978), whereas the presence of Sil–St in the eastern Quetico is diagnostic of bathozones 3 and 4 (3.4–5.5 kbar).

As noted by Williams (1991), tectonic thickening of the sedimentary pile and intrusion of minor I-type granitic rocks occurred prior to the thermal acme. Most of the large pre- to syntectonic granitic bodies are peraluminous and have sedimentary sources but display little evidence of thermal contact metamorphism; one exception is the South Beatty Lake pluton in the northern Quetico subprovince (Pirie & Mackasey 1978). Steeply dipping thermal gradients, local increases in temperature around large plutons, and the general association of the highest-grade rocks with abundant generation of leucosome, indicate that the source of heat was a combination of burial, upward magmatic transport, and tectonism.

In the northern Quetico, M_1 metamorphism is estimated to have occurred between 2698 Ma, the maximum age of sedimentation (Davis *et al.* 1990) and 2688 ± 4 Ma, the age of emplacement of the late syntectonic Blalock pluton (Davis *et al.* 1990). In the southern Quetico, M_1 occurred after 2690 Ma, the maximum age of sedimentation (Zaleski *et al.* 1999). The timing of M_2 metamorphism is more poorly constrained, and may have been protracted. In the Manitouwadge area, Zaleski *et al.* (1999) constrained regional D_2 deformation to 2680–2677 Ma, and suggested that migmatization in both the northern Wawa and southern Quetico subprovinces occurred after 2679 Ma, broadly coincident with D_3 deformation. This inference is consistent with observations elsewhere in the Quetico that contact aureoles around late plutons, dated at 2671 ± 2 to 2665 ± 2 Ma, as well as late granitic pegmatites dated at 2653 ± 4 Ma, overprint regional metamorphic fabrics (Percival & Sullivan 1988, Percival 1989). North of Manitouwadge, Pan *et al.* (1998) reported a U–Pb zircon age of 2666 ± 1 Ma from a granitic pegmatite concordant with respect to the D_3 fabric, and suggested that the regional amphibolite-facies metamorphism occurred between 2671 and 2665 Ma, consistent with the ages cited above. The timing of peak granulite-facies metamorphism north of Manitouwadge appears to be some 15 Ma younger, on the basis of U–Pb zircon ages of 2650 ± 2 and 2651 ± 3 Ma from a mafic granulite and a tonalitic leucosome, respectively (Pan *et al.* 1998). Zaleski & van Breemen (1997) reported that titanite ages young with increasing metamorphic grade, ranging from ~2686 Ma in the southern Manitouwadge greenstone belt to ~2640 Ma in the southern Quetico, suggesting that the thermal effects of regional metamorphism may have lasted over ~30 million years, from 2677 to 2640 Ma, in the higher-grade parts of the Wawa and Quetico subprovinces. On the basis of their regional geological and geochronological studies, Zaleski *et al.* (1999) concluded that M_2 metamorphism occurred after the tectonic juxtaposition of the Quetico and Wawa subprovinces.

Wawa Subprovince

Williams *et al.* (1991) asserted that the large variation in regional metamorphic grade present within the Wawa subprovince is related, in part, to the ratio of supracrustal to plutonic rocks; the lowest grades of metamorphism are found within large greenstone belts that contain few internal granitic bodies (*e.g.*, Michipicoten greenstone belt), and the highest grades of metamorphism are found in small greenstone belts containing a greater volume of internal granitic rocks (*e.g.*, Manitouwadge greenstone belt). In addition, metamorphic grade increases in the Wawa suprovince from Lake Superior eastward toward the Kapuskasing structural zone, generally reflecting increasing depths of exposure. In the Manitouwadge area, metamorphic grade increases northward toward the Quetico subprovince, as discussed in the previous section.

Major greenstone belts subjected to low-grade metamorphism include the Mishibishu, Michipicoten and Gamitagama belts in the east, and the Shebandowan and Vermillion belts in the west. All exhibit a metamorphic zonation similar to that observed in other granite–greenstone subprovinces, with greenschist-facies cores and amphibolite-facies zones developed marginal to large internal and external plutons (*e.g.*, Ayres 1969, Studemeister 1983, 1985, McGill & Shrady 1986). In the Wawa area, Studemeister (1983, 1985) suggested that pretectonic tonalite stocks produced greenschist-facies contact aureoles at 450–550°C and <2 kbar, with regional metamorphism affecting both the volcanic and plutonic rocks under conditions of 325–450°C and 2–3 kbar. In both the Shebandowan and Vermillion greenstone belts, regional metamorphism was associated with D_2 deformation, which affected Timiskaming-type metasedimentary and metavolcanic rocks. The timing of deformation and metamorphism is bracketed at 2685–2680 Ma (Corfu & Stott 1986, 1998). Although 2890–2880 Ma supracrustal and plutonic rocks have been reported from the Hawk Lake area of the Michipicoten greenstone belt (Turek *et al.* 1984, 1992), there is no indication of Geon 28 regional metamorphism in the Wawa subprovince.

Major greenstone belts subjected to high-grade metamorphism include the Hemlo, Schreiber and Manitouwadge belts in the north-central and northeastern Wawa subprovince. The Manitouwadge greenstone belt was subjected to upper-amphibolite-facies metamorphism (600–650°C, 3.0–5.0 kbar, James *et al.* 1978, Robinson 1979, Pan & Fleet 1992), at 2675 ± 3 Ma to 2671 ± 3 Ma, based on U–Pb monazite ages (Davis *et al.* 1994, Zaleski & van Breemen 1997). Kyanite has been locally reported within the Hemlo belt and has been the focus of considerable study (see summary in Muir 1997). Largely on the basis of the paragenesis of the aluminosilicate phases, there is evidence of a polymetamorphic history in the Hemlo belt (Powell *et al.* 1999, Pan & Fleet 1993). This history consists of (1) contact metamorphism (2–4.5 kbar) that produced early-formed andalusite, related to the 2688–2684 Ma (Corfu & Muir 1989) Heron Bay – Cedar Lake intrusive suite, (2) early amphibolite-facies metamorphism that produced kyanite (600–650°C, 6–7 kbar, bathozone 5) associated with D_1 at 2688–2684 Ma, and (3) a dominant regional metamorphism associated with the widespread development of sillimanite. The latter event produced greenschist–amphibolite-facies conditions in the west (510–530°C, 3–3.5 kbar, bathozone 3: Powell *et al.* 1999), increasing to amphibolite facies in the east (550–600°C, 4–4.5 kbar, bathozone 4: Powell *et al.* 1999). This regional metamorphism is considered to be associated with D_3 (Corfu & Muir 1989, Pan & Fleet 1993) and occurred at ~2678 Ma. Muir (1997) noted two morphological and textural varieties of garnet locally in the Hemlo belt, as well as the presence of retrograded lenses of cordierite, consistent with a polymetamorphic history and the bathozone-3 to -4 conditions associated with the younger regional metamorphic event.

The Wawa gneiss domain lies between the Michipicoten–Gamitagama greenstone belts and the Kapuskasing structural zone. Geobarometric studies of amphibole-bearing tonalitic rocks across the Wawa gneiss domain indicate pressures of crystallization of 5 kbar in the west, increasing to 6.5 kbar in the east (Percival 1990). U–Pb ages obtained on titanite also decrease from 2685 Ma in the west to ~2600 Ma in the east, possible reflecting prolonged high temperatures at deeper structural levels for rocks in the eastern part of the domain (Krogh & Moser 1994, Percival & West 1994).

Metamorphism and deformation in the Wawa subprovince have been related to accretion to the Quetico subprovince (*e.g.*, Williams *et al.* 1991, Card & Poulsen 1998). There is a similarity of timing and sequence of events between the two subprovinces, with evidence of an earlier, localized, amphibolite-facies metamorphic event in both occurring at between 2700 and 2688 Ma, followed by a regional, slightly lower-pressure event at 2689–2678 Ma.

Kapuskasing Structural Zone

The Kapuskasing structural zone represents a section through the crust, extending from low-grade supracrustal rocks of the Michipicoten greenstone belt corresponding to depths of 5 to 10 km, to granulite-facies rocks corresponding to depths of 25 to 30 km. The section exposes three main layers: 1) a 10–15 km upper layer dominated by low-grade supracrustal rocks and discordant plutons, 2) a 10–20-km-thick middle layer consisting of gneissic and plutonic rocks, and 3) a lower layer of heterogeneous granulite- to upper-amphibolite-facies gneiss (Percival & West 1994). The Kapuskasing structural zone transects the easterly-trending metasedimentary and granite–greenstone belts in the Superior Province, and can be divided into several domains on the basis of geological and geophysical attributes (Percival & McGrath 1986). From north to south, these are the Fraserdale–Moosonee, Groundhog River and Chapleau blocks, as well as the Val Rita block in the Wawa gneiss domain of the Wawa subprovince. The geology, geophysics and tectonic history of the Kapuskasing structural zone have been summarized by Percival & West (1994), from which this account is adapted.

The Fraserdale–Moosonee block consists mainly of granulite-facies paragneiss and mafic gneiss with evidence of 8 to 10 kbar pressures (Percival & McGrath 1986), whereas adjacent granulite-facies paragneiss in the Quetico subprovince to the west yield pressures of 5–6 kbar (Percival & McGrath 1986), suggesting ~13 km of uplift of the Fraserdale–Moosonee block relative to the adjacent Quetico subprovince. The Groundhog River block consists of granulite-facies mafic and tonalite gneiss, with evidence of 7–9 kbar pressures (Leclair 1992), and metamorphic zircon with ages of 2657 Ma and 2648 Ma (Leclair & Sullivan 1991), similar to those present in the Chapleau block.

Mafic gneiss and paragneiss in the Chapleau block are considered to be of supracrustal origin, similar in age to supracrustal rocks in the adjacent Abitibi and Wawa subprovinces (Shaw *et al.* 1994). The southwestern boundary of the block is a complex zone of lithological, structural, and metamorphic transition from granulite-facies rocks to amphibolite-facies gneisses and supracrustal rocks of the Wawa subprovince (Percival 1983). Across this transition, metamorphic pressures decrease from 8–11 kbar in the granulites of the Kapuskasing structural zone to 5–6 kbar in the Wawa gneiss domain (Percival & West 1994). Thermometry on the orthopyroxene-bearing gneisses in the Chapleau block indicates temperatures of 720–780°C (Percival & West 1994). Hartel & Pattison (1996) suggested that peak temperatures and pressures based on garnet – diopside – plagioclase – quartz thermobarometry in the Shawmere area of the Chapleau block, rather than being at 9 kbar and 685–735°C, are actually closer to 11 kbar and 850°C if one takes into account studies of dehydration-melting of amphibolite (see also Mäder *et al.* 1994). Consequently, peak thermobarometric measurements in the Kapuskasing structural zone should be regarded as minimum estimates.

Zircon grains from mafic gneiss, considered to date granulite-facies metamorphism, give a range of ages, mainly between 2663 and 2630 Ma (Krogh 1993). Zircon ages appear to decrease with increasing paleodepth from 2663 Ma in the west to complex populations with components of 2650, 2645, 2636, 2630, 2616 and 2582 Ma in the east (Krogh 1993, Krogh & Moser 1994). Regional thermobarometry and geochronology show that granulite units of different ages formed at similar structural levels, suggesting either polychronous metamorphism, or a model in which the zircon age patterns reflect complex interactions among temperature, strain, fluid flow and fluid composition (Card & Poulsen 1998). U–Pb titanite (2642 Ma) and monazite (2654, 2649 Ma) ages from the upper-amphibolite Agawa migmatite domain to the south (Corfu 1987) suggest a similar early history to that recorded in the Chapleau block, even though the Chapleau block contains titanite and monazite as young as 2500 and 2510 Ma, respectively (Krogh & Moser 1994). Ar–Ar hornblende ages indicate that the Wawa gneiss domain cooled through 500°C at ~2570 Ma, whereas rocks in the Shawmere area of the Chapleau block closed at ~2470 Ma (Lopez Martinez & York 1990, Hanes *et al.* 1994).

U–Pb zircon geochronology in the Kapuskasing structural zone indicates protracted metamorphism and mineral growth in the interval 2695 to 2584 Ma, with later resetting of titanite (U–Pb, 2504 Ma), hornblende (K–Ar, 2480 Ma), and biotite (K–Ar 2000 Ma, Rb–Sr 1930 Ma) (Percival & West 1994 and references therein). The Kapuskasing rocks were probably uplifted during several events, some older than the 2477 Ma Matachewan dike swarm, some younger (Percival & West 1994, Halls *et al.* 1994, Halls & Zhang 1998). The ~2040 Ma Kapuskasing dikes were emplaced prior to movements on boundary faults to the Kapuskasing structural zone (Percival *et al.* 1994), but the 1888 Ma Cargill complex is only slightly offset by these faults.

CENRAL ABITIBI SUBPROVINCE

The central Abitibi subprovince is a granite–greenstone terrane consisting mainly of the Swayze and Abitibi greenstone belts. In contrast, the southwestern Abitibi subprovince is a granitic gneiss – plutonic domain, termed the Ramsey–Algoma granitoid complex (Jackson & Fyon 1991), which contains a few small greenstone belts. The central Abitibi is the better studied of the two; most descriptions of the Abibiti subprovince focus on the Abitibi greenstone belt.

Two main metamorphic episodes have been described for the central Abitibi subprovince. The M_1 event, in some cases referred to as pre-Kenoran, consists of seafloor metamorphism of the metavolcanic rocks to produce Prh–Pmp assemblages (Dimroth & Lichtblau 1979, Dimroth *et al.* 1983, Jolly 1978, 1980). M_1 was accompanied by localized zones of hydrothermal alteration. In addition, pre-Kenoran synvolcanic plutons, such as the Lac Dufault, Flavrian and Powell plutons (2690 ± 2, 2701 ± 2 Ma; respectively, Mortensen 1993) display contact-metamorphic aureoles. The early seafloor metamorphism is not associated with cleavage development (Dimroth *et al.* 1983).

M_2, the regional "Kenoran" event, varies from subgreenschist facies in the core of the Abitibi belt to greenschist facies to the north, south and west (Jolly 1974, 1977, 1978, 1980, Powell *et al.* 1990). Generally narrow (<5 km wide) amphibolite-facies contact aureoles are well developed near the late (~2680 Ma) alkaline intrusions and near most of the large tonalite–granodiorite bodies (*e.g.*, 2689 ± 1 Ma Lake Abitibi batholith: Mortensen 1993). M_2 regional greenschist-facies metamorphism affects the large tonalite–granodiorite bodies older than ~2690 Ma to varying degrees, with the intensity of the metamorphism being related to degree of fracturing (Jolly 1978, Goldie 1979). The greenschist- and amphibolite-facies assemblages spatially associated with the granitic intrusions apparently overprint and postdate prehnite–pumpellyite-facies assemblages (Jolly 1978, 1980, Dimroth & Lichtblau 1979), although the textural features cited also could be explained by outward progradation of the thermal anomaly associated with the emplacement of the granitic rocks (Jackson & Fyon 1991).

Metamorphic facies reported by Jolly (1980) are, in part, limited by faults. An example is the Mulven fault, where subgreenschist-facies assemblages are recorded northwest of the fault, and greenschist–subgreenschist-facies, to the southeast. Powell *et al.* (1990) suggested that M_2 regional metamorphism predates at least some displacement along the Larder–Cadillac deformation zone. On a regional scale, higher metamorphic grades (greenschist to amphibolite) are found south of the Larder–Cadillac deformation zone than to the north (greenschist to subgreenschist). In addition, metamorphic aureoles around pre-2690 Ma granitic bodies and ~2680 Ma alkaline intrusions (*e.g.*, Lebel stock) predate the Larder–Cadillac deformation zone.

The conditions of M_2 regional metamorphism are not well understood for most of the Abitibi greenstone belt. The well-developed prehnite–pumpellyite-facies zone suggests a maximum temperature of approximately 300°C and pressures between 2 and 3 kbar (see P–T grid of Liou *et al.* 1987). Detailed thermobarometry in the Matachewan area by Powell *et al.* (1993) suggests that the M_2 greenschist-facies metamorphism occurred at 250–270°C and 2–2.5 kbar. In Quebec, Crd–Ath and andalusite-bearing assemblages in metamorphosed altered metavolcanic rocks present in the contact aureole of the Lac Dufault granodiorite suggest a pressure of metamorphism less than 3.4 kbar (bathozone 2) at the time of intrusion. Most of the region south of the Larder–Cadillac zone of deformation in Ontario is at the greenschist facies, typically Ep–Act-bearing, indicating minimum temperatures of metamorphism of 300–350°C.

M_1 likely occurred between 2700 Ma (age of the youngest Keewatin-type metavolcanic rocks) and prior to ~2690 Ma, the age of the younger tonalite–granodiorite batholiths. M_2 likely occurred at ~2670 Ma, and affected ~2680 Ma Timiskaming metasedimentary and related rocks (Corfu 1993). A titanite age of 2665 ± 4 Ma from Kirkland Lake (Wilkinson *et al.* 1999) and a monazite age of 2659 ± 3 Ma from the Timmins area (Davis *et al.* 1994) provide minimum ages for peak M_2 metamorphism in the Abitibi subprovince.

Most Ar–Ar studies in the central Abitibi subprovince have focussed on the timing of gold mineralization and shear-zone development (*e.g.*, Masliwec *et al.* 1986, Hanes *et al.* 1992, Zweng *et al.* 1993, Powell *et al.* 1995) rather than cooling history. Data presented by Feng *et al.* (1992) for the 2698 ± 4 Ma Round Lake batholith (U–Pb Zrn, Mortensen 1993) indicate hornblende closure at 2669 ± 6 Ma. In contrast, Powell *et al.* (1995) reported excess Ar signatures from the hornblende grains present in weakly strained rocks that they studied. Muscovite ages from the Preissac–Lacorne batholith are 2615–2594 Ma (Feng *et al.* 1992); again, Powell *et al.* (1995) reported a large variation in muscovite ages from weakly strained Abitibi rocks, ranging from 2594 to 2509 Ma, with biotite showing a similar pattern, with a range of 2422–2504 Ma. These data suggest that despite the low metamorphic grade of the Abibiti subprovince, several events involving low-grade hydrothermal alteration occurred subsequent to regional metamorphism, making determination of a reliable pattern of regional cooling problematic.

Although many investigators have suggested that the Abitibi and Wawa subprovinces may be contiguous (see Williams *et al.* 1991, Card & Poulsen 1998), the metamorphic conditions and timing of regional metamorphism within the whole Abitibi subprovince are distinct from those observed in the lower-grade parts of the Wawa subprovince (Table 1), suggesting that a more complex relationship exists between the Abitibi and Wawa subprovinces.

Southwestern Abitibi Subprovince

Greenstone belts in the southwestern Abitibi subprovince are dominantly at the amphibolite facies, but greenschist-facies cores are found in the larger belts (*e.g.*, Batchawana, Temagami). In the Batchawana belt, metamorphism occurred between 2674 ± 3 Ma and 2668 ± 2 Ma, on the basis of a metamorphic titanite age and the age of a postmetamorphic pluton (Corfu & Grunsky 1987). This age range is similar to that for M_2 metamorphism within the central Abitibi subprovince.

The Ramsey–Algoma granitoid complex, which forms much of the southwestern Abitibi subprovince, includes granodioritic gneisses emplaced between 2716 and 2711 Ma (Krogh *et al.* 1984, Corfu & Grunsky 1987, Heather & van Breemen 1994); these rocks are regarded as subvolcanic intrusions coeval with volcanism throughout the Abitibi subprovince. The gneisses are cut by younger granitic plutons emplaced between 2692 and 2642 Ma (*e.g.*, Card & Poulsen 1998, Ames *et al.* 1997). This age range overlaps, and extends for some 20 million years after, the range of major plutonism in the central Abitibi.

The metamorphic history of the southwestern Abitibi subprovince is not well known. In Figure 1, areas underlain dominantly by gneissic rocks of the Ramsey–Algoma granitoid complex are shown as having undergone amphibolite-facies metamorphism, with the areas of plutonic rocks shown as syn- to postmetamorphic. An exception is the granulite-facies Levack gneiss complex located between the 1850 Ma Sudbury Igneous Complex and the Benny greenstone belt. The Levack gneiss complex contains metaplutonic rocks ranging in age from 2711 to 2647 Ma (Ames *et al.* 1997), with granulite-facies metamorphism at 6–8 kbar and 750–800°C (James *et al.* 1992) occurring between 2647 and 2642 Ma (Ames *et al.* 1997). James *et al.* (1992) also noted local amphibolite-facies retrogression of the gneisses to 2–3 kbar and 500–550°C adjacent to the 2642 ± 1 Ma postmetamorphism Cartier granite (Meldrum *et al.* 1997). Within 2 km of the Sudbury Igneous Complex, the granulites were overprinted by a second metamorphic event at 1850 Ma, with conditions of 4–6 kbar and 800–950°C reported (James *et al.* 1992). On the basis of a U–Pb and Ar–Ar study of the Levack gneiss complex, Ames *et al*. (1997) suggested a two-stage exhumation history of the granulite-facies rocks: a late Neoarchean uplift from ~18–24 km to ~6–9 km associated with emplacement of the Cartier granite, with final unroofing occurring during the 1850 Ma Sudbury event.

Ames *et al.* (1997) noted that the timing of plutonic and metamorphic events within the Levack gneiss complex is similar to that observed in the Kapuskasing structural zone. What differs between the two areas is their exhumation history, with the Levack gneiss being uplifted to higher crustal levels very soon after peak metamorphism, possibly owing to crustal extension, whereas uplift of the Kapuskasing structural zone occurred much later, perhaps in the Paleoproterozoic, along a listric thrust fault (*e.g.*, Percival & West 1994).

Pontiac Subprovince

Only limited exposures of the Pontiac subprovince are found in Ontario, as these rocks are mainly covered by Paleoproterozoic rocks of the Southern Province. Probable metamorphosed equivalents of the Pontiac subprovince, known as the Red Cedar Lake gneiss, occur in the Grenville Front tectonic zone south of Temagami (*e.g.,* Easton 1992). The Pontiac subprovince was metamorphosed to much higher grades than rocks of the adjacent Abitibi greenstone belt (Jolly 1978, 1980), with the presence of St, Sil, Ky, and partial melting indicating temperatures above 550°C, and pressures suggestive of bathozone 3 or 4. A U–Pb titanite age of 2660 ± 6 Ma from orthogneiss within the Pontiac subprovince in Quebec (Machado *et al.* 1991) gives some indication of the timing of regional metamorphism.

Overview and Discussion

Relationship of metamorphic domains to geophysical data

The aeromagnetic and gravity signature of high-grade metamorphic rocks in Ontario is distinct. Aeromagnetic data show either a birds-eye-maple pattern, or a pattern of striped, linear magnetic highs (Gupta 1991), with the change in magnetic character of the rocks generally corresponding to the location of the orthopyroxene-in isograd. In the English River subprovince, Thurston & Breaks (1978) noted a correspondence between high-grade rocks, a birds-eye magnetic pattern, and a Bouguer gravity high. During the course of this study, it was observed that this association of magnetic and gravity anomalies is present for both low- and intermediate-pressure granulite domains, and that the presence of paired gravity highs on Bouguer gravity and vertical-gravity-gradient maps (Gupta 1992a, b) is a more reliable indicator of granulite-facies metamorphic rocks than Bouguer gravity highs alone. The paired gravity anomalies seem to be related to the appearance of the assemblage Crd–Grt–Kfs, and as such, they define areas of upper-amphibolite- to granulite-facies rocks in metasedimentary gneisses, unlike the magnetic signature that is indicative solely of granulite-facies domains.

Regional metamorphic patterns and synchronism of metamorphic events during Geon 26–27

Card (1990), Williams *et al.* (1992) and Card & Poulsen (1998, Fig. 2.91) suggested that within the granite–greenstone subprovinces of the Superior Province, there was a general southward younging of metamorphism, from ~2720–2700 Ma in the north, through ~2710–2690 in the central Superior, to 2695–2685 Ma in the south. Card & Poulsen (1998) noted that this trend reflects a similar younging trend in the ages of periods of major volcanism, plutonism and deformation within the Superior Province, and that this progressive younging is consistent with southward coalescence of the Superior Province during the Neoarchean. Although this basic thesis may be correct with respect to the tectonic development of the Superior Province, the pattern and timing of metamorphism within the Superior Province are not straightforward.

The regional metamorphic pattern in the Superior Province after ~2710 Ma largely reflects the distribution of subprovince types and exposed crustal level, as follows: 1) granite–greenstone and plutonic subprovinces generally show peak metamorphism at ~2700 Ma. The Abitibi is a notable exception to this pattern, in part because it shows evidence for two main pulses of metamorphism. 2) Metasedimentary and high-grade gneiss subprovinces show evidence for protracted metamorphism, with an earlier event at 2693–2680 Ma, and a second event at 2678–2665 Ma. In areas of granulite-facies rocks, regardless of pressure regime, U–Pb isotopic systems may record ages as young as 2640–2630 Ma.

At least five major metamorphic events can be outlined in the period 2710 to 2640 Ma; these generally correspond to major episodes of plutonism within the Superior Province (Fig. 4, Table 2). For the purposes of discussion, these events are designated P_1 through P_5, to avoid confusion with previously labeled metamorphic (*M*) events in the text. In the lowest-grade areas, P_1 is typically the only regional event; with increasing metamorphic grade or increasing depth in the crust, or both, however, additional events are typically recorded. In addition, two phases of granulite-facies metamorphism appear to be present in most granulite-facies terranes, an older, syndeformational event (P_2 ~2693–2680 Ma) and a younger, thermal event (P_4 ~2660–2648 Ma). Evidence for earlier (pre-2710 Ma) metamorphic events is discussed elsewhere in this paper.

P_1 occurs between 2710 and 2695 Ma, and is best preserved in the granite–greenstone subprovinces (Fig. 4), where in many areas, it is the main regional metamorphic event (*e.g.*, Uchi, Wabigoon). In the Wawa subprovince, P_1 is predominant in the lower-grade greenstone belts (Mishibishu, Michipicoten, Gamitagama, Shebandowan, Vermillion), P_2 in the higher-grade belts (Hemlo, Schreiber, Manitouwadge). In both the Abitibi and Wawa subprovinces, a second regional event (P_3 ~2680–2665 Ma) is recorded. In the Quetico metasedimentary subprovince, P_1 is locally preserved along the northern margin of the subprovince, with P_3 being the main event. P_1 may be 5 to 10 Ma younger in the Abitibi and Wawa subprovinces (~2700–2690 Ma) than in the Wabigoon and Uchi subprovinces (~2700–2705 Ma). This age break seems to be an abrupt transition, rather than a gradational southward younging.

P_2 (~2693–2680 Ma) occurs mainly in the metasedimentary, plutonic and high-grade subprovinces (English River, Winnipeg River, Quetico), and in the higher-grade parts of the Wawa subprovince. In all these areas, P_2 is associated with major regional deformation, and locally attains the granulite facies. P_3 (~2679–2665 Ma) is present in the Abitibi, English River, the higher-grade parts of the Quetico and Wawa subprovinces and the Kapuskasing structural zone. In the southern Superior Province, it is the dominant regional metamorphic event.

P_4 (~2660–2648 Ma) is only present as a significant regional event in the granulite-facies domains of the Quetico, southern Abitibi, English River, and Pikwitonei subprovinces and within the Kapuskasing structure. However, plutons of this age occur scattered throughout the Superior Province, and many of the S-type granites and associated rare-element pegmatites present within the Superior Province are of this age (see also Percival & Skulski 2000). Moser *et al.* (1996) suggested that within the Wawa gneiss domain and the Kapuskasing structural zone, there were two discrete events within P_4, one at ~2660 Ma, the other at ~2645 Ma. P_5 (<2640 Ma) is only present as a regional event within the deepest structural levels revealed in the high-grade domains within the Kapuskasing structural zone and the Pikwitonei subprovince.

The tectonic significance of the above pattern in metamorphic ages and association with level in the crust is not well understood, and it would be incorrect to interpret the pattern as simply one of metamorphic younging with depth. One reason is that the distribution of reliable zircon and titanite ages of metamorphism varies among subprovinces. Thus the apparent complexity of metamorphic ages in the southern Superior Province, for example, may simply reflect the fact that the area is better studied. Another reason is that for much of the Superior Province, the relationship between major magmatic events and regional metamorphism is unknown. It is not clear, for example, if 1) mantle-derived heat, in the form of plutons, drove metamorphism, 2) loading and thermal relaxation of the crust caused crustal melting and associated metamorphism, 3) crustal underplating and delamination drove metamorphism, or 4) some combination of the above. Addressing the question of heat source is further complicated if the Archean mantle and crust were hotter than at present. Moser *et al.* (1996) proposed a model of arc–continent collision and crustal delamination and extension that explains the observed timing and distribution of metamorphism within the southern Superior Province, but whether this model can be applied elsewhere in the Superior Province remains to be demonstrated.

TABLE 2. METAMORPHIC PRESSURES AND DISTRIBUTION OF NEOARCHEAN METAMORPHIC EVENTS IN THE SUPERIOR PROVINCE

Metamorphic Event	Bathozone (depth in the crust) 1–2 (3–8 km)	3–4 (9–14 km)	5–6 (15–>24 km)
P_1 (2710–2695 Ma)	northwest Superior, Uchi, Wabigoon, central Abitibi	southern Winnipeg River	Quetico
P_2 (2693–2680 Ma)	Wawa	English River, northern Winnipeg River	Kapuskasing, Pikwitonei
P_3 (2678–2665 Ma)	western Quetico, central Abitibi	southwestern Abitibi, English River, northern Winnipeg River, Wawa (high-grade), central and eastern Quetico	south-central Quetico
P_4 (2660–2648 Ma)		northern Wawa, western Wawa gneiss domain	south-central Quetico granulites, Kapuskasing, eastern Wawa gneiss domain, Levack granulites
P_5 (<2640 Ma)			eastern Kapuskasing, western Pikwitonei

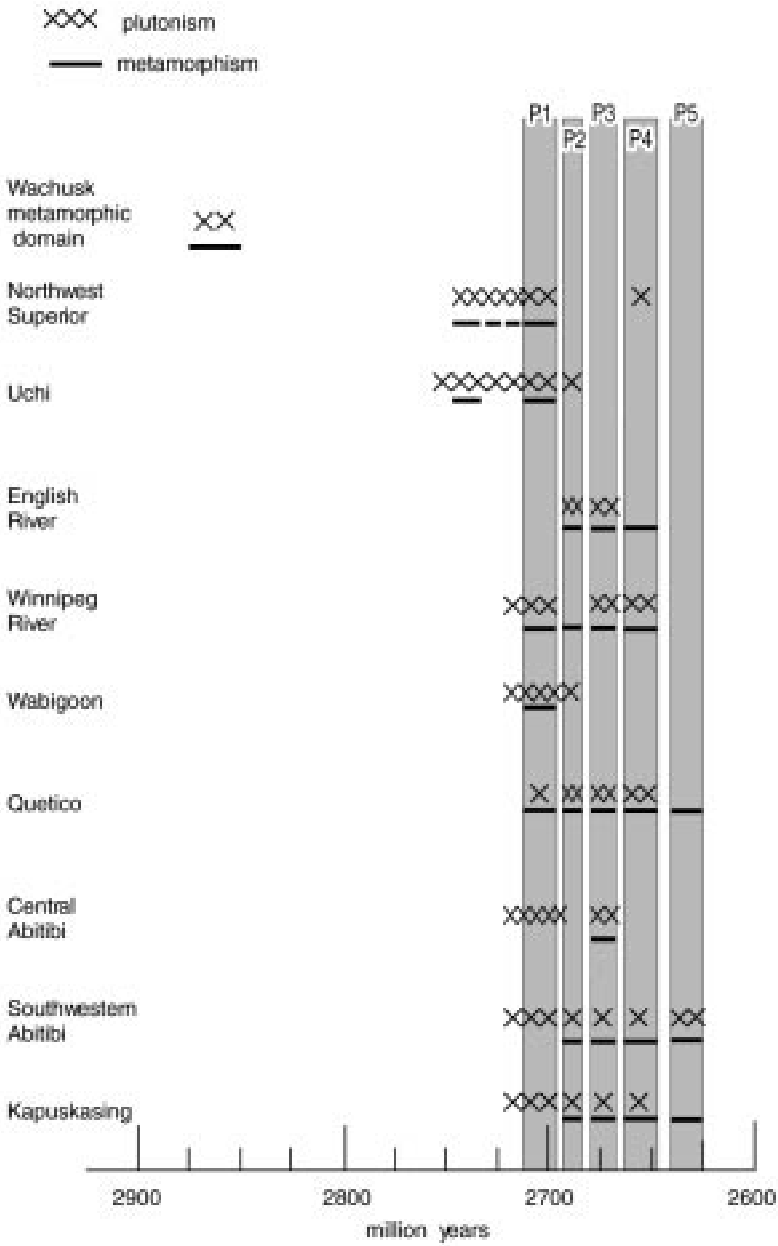


FIG. 4. Summary of geochronological data for Archean metamorphism and plutonism in the Superior Province.

Relationship between Mesoarchean crust and pre-2800 Ma metamorphic events

Figure 5 outlines the distribution of areas of pre-2800 Ma crust and supracrustal sequences in the Superior Province as well as areas where pre-2800 Ma metamorphic events can be documented. The area of known pre-2800 metamorphism is considerably smaller than the possible distribution of older crust. This fact reflects three factors: 1) much of the region in question is intruded by 2740–2700 Ma felsic plutons, with only small remnants of older crust preserved, 2) intensive metamor-

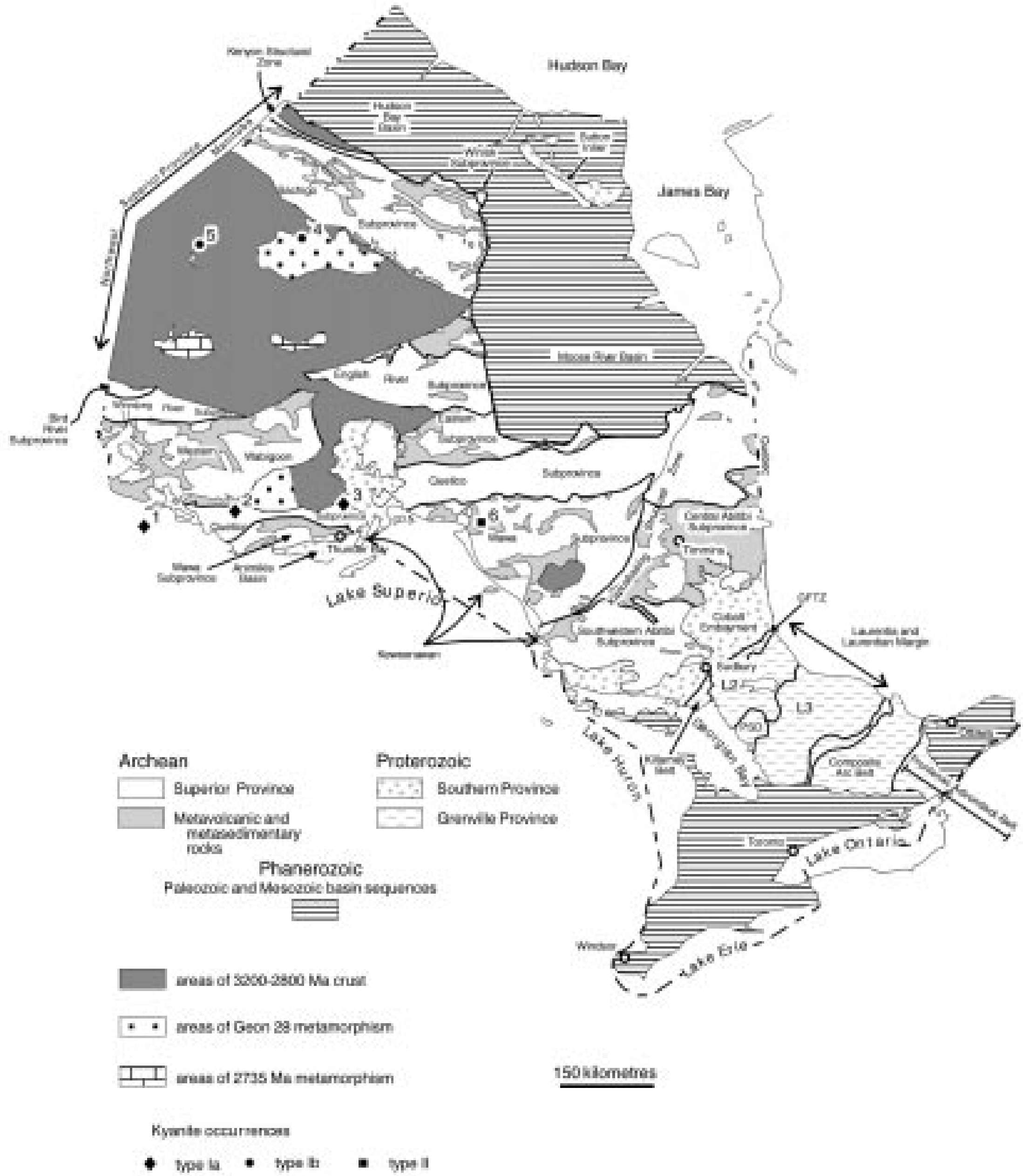


FIG. 5. Distribution of pre-2800 Ma crust, Geon 28 metamorphic events, and kyanite occurrences within the Superior Province of Ontario. Kyanite types after Pan & Fleet (1999).

phism at ~2700 Ma has largely obscured evidence of older metamorphism, and 3) there is a paucity of precise ages from much of the northwestern Superior Province.

In addition, the present distribution of older crust likely does not reflect its original distribution in the Mesoarchean and Early Neoarchean. Tomlinson *et al.* (1999) have suggested, on the basis of stratigraphy, and age and type of volcanism and plutonism, that two main Mesoarchean crustal blocks existed in the western Superior Province: 1) a North Caribou – south-central

Wabigoon block, that experienced felsic magmatism between 3020 and 2960 Ma, followed by deposition of quartz arenite sequences and subsequent 2950–2920 Ma mafic volcanism, and 2) a north-central Wabigoon – Winnipeg River block, containing older crust (3000–3170 Ma) that was also affected by 2950–2920 Ma mafic magmatism. To date, Geon 28 metamorphic rocks are found only in the North Caribou – south-central Wabigoon block. Further studies of Mesoarchean gneisses in western Superior are needed to evaluate the extent and metamorphic history of this older crust.

Regional variations in pressure

Figure 6 summarizes regional variations in pressure, expressed as bathozone levels, for the 2700–2660 Ma metamorphism present within the Superior Province. The observed pattern largely reflects uplift in the late Neoarchean and the Proterozoic. Granite–greenstone subprovinces generally show the least postmetamorphic uplift, high-grade gneiss subprovinces generally showing the most. Only in the Quetico subprovince is an east–west variation in relative uplift well documented (Fig. 6). Moser *et al.* (1996) argued that the limited variation in level of the crust exposed and range in metamorphic grade in the Superior Province are lower than what would be expected following Himalayan or Alpine-style collision. They attributed development of the observed pattern to a period of cyclic collision and delamination of structurally underplated material at the base of the juvenile Superior craton between 2670 and 2640 Ma, followed by differential uplift during the Archean and Proterozoic.

Tectonic significance of high-pressure minerals

It is worth noting a few metamorphic anomalies within the Superior Province. Early-formed kyanite occurs in the North Caribou Lake greenstone belt, close to the fault boundary of the North Caribou terrane with a younger crustal block to the north. Pan & Fleet (1999) attributed the development of this early kyanite to higher metamorphic pressures caused by crustal thickening during tectonic accretion of arc and continental margin sequences in the Superior Province. If so, this event must have occurred prior to ~2870 Ma, the age of the contact aureole of the North Caribou Lake batholith, much earlier than that associated with M_1 (P_1) metamorphism along the Quetico–Wabigoon subprovince boundary to the south that Pan & Fleet (1999) also related to crustal thickening during accretion (Fig. 5). Kyanite in both the Favourable Lake and North Caribou Lake belts occurs near major regional faults, suggesting, however, that kyanite formed during Kenoran crustal thickening. Along the southern boundary of the North Caribou – south-central Wabigoon block, there is local evidence for preservation of ~2735 Ma and ~2715 Ma metamorphism that may be related to the accretionary history of this margin, as outlined by Stott & Corfu (1991). An early higher-pressure event has also been documented along the Winnipeg River – Wabigoon subprovince boundary at ~2685 Ma (Menard *et al.* 1997), which was attributed to crustal thickening during accretion. In the Quetico subprovince, evidence for an early 2702–2688 Ma kyanite-forming metamorphic event is reported for several sites along the Quetico–Wabigoon subprovince boundary (Fig. 5). Pan & Fleet (1999) suggested that the kyanite occurrences along the Quetico–Wabigoon subprovince boundary result from higher metamorphic pressures caused by crustal thickening during tectonic accretion.

Relevance of Ar–Ar ages to cooling and uplift history

The distribution of Ar–Ar hornblende and biotite ages for the Superior Province are summarized in Table 3, and show a general correspondence to the overall regional metamorphic pattern, with the structurally highest granite–greenstone subprovinces (Uchi, central Abitibi) exhibiting relatively rapid cooling histories, and with the structurally lowest metasedimentary and high-grade gneiss subprovinces exhibiting complex cooling histories (Wawa gneiss domain, Kapuskasing), commonly complicated by the widespread occurrence of excess argon within hornblende and biotite. Hanes & Archibald (1999) noted an east-to-west variation in Ar–Ar hornblende and biotite ages across the English River and Winnipeg River subprovinces (Table 3), which they attributed to west-side-up and north-side-up tilting at, or after, ~2400 Ma. Table 3 does not include the sporadic occurrence of younger (2300–1800 Ma) biotite ages adjacent to major east–west faults throughout the Superior Province (*e.g.*, Easton 1985, Hanes & Archibald 1999), which may reflect reactivation of these structures during the Paleoproterozoic. Along the northern margin of the northwest Superior Province, K–Ar and Ar–Ar ages of 1800 Ma are likely related to distal effects of the Trans-Hudson orogen to the north.

Synthesis

The observed distribution of metamorphism in the Superior Province outlined above is broadly consistent with the evolution of the Superior Province outlined by Williams *et al.* (1992) and Card & Poulsen (1998). According to these models, following a complex Meso- and Early Neoarchean history, a complex island arc, back-arc, oceanic plateaus and microcontinent system developed between 2740 and 2720 Ma. Through a series of complex interaction of plates, amalgamation of greenstone belts and older cratonic elements occurred over the interval 2740–2690 Ma, likely largely thin-skinned in character. Amalgamation appears to have started earlier in the north, and migrated progressively southward. Metamorphic events related to this early history of assembly are locally preserved (P_1). Amalgamation was

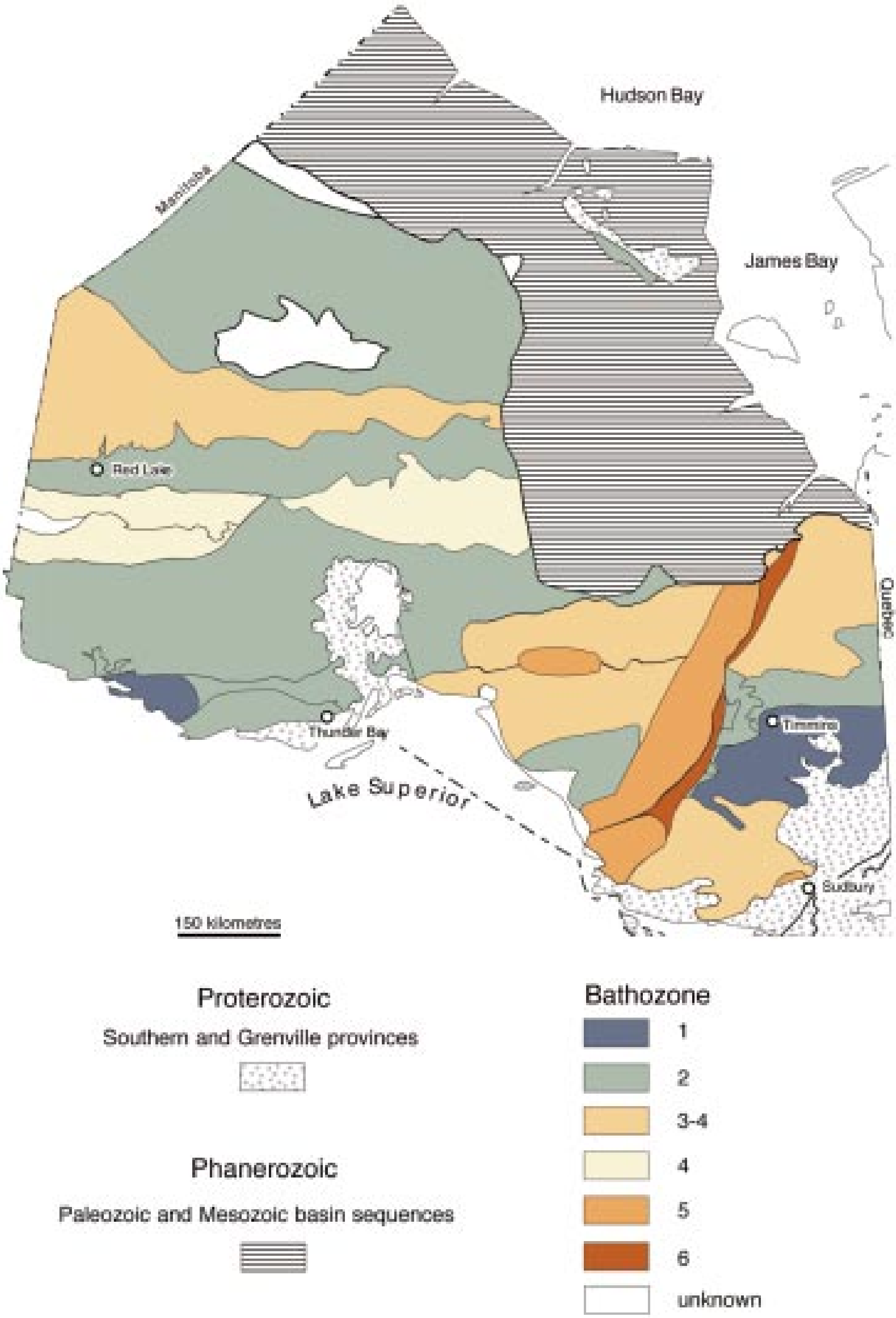


FIG. 6. Distribution of regional variations in pressure associated with 2700–2660 Ma metamorphism within the Superior Province on a subprovince scale, expressed in terms of bathozones. The map was produced using a combination of mineral-assemblage data and thermobarometric data. Note that along many subprovince boundaries (*e.g.*, Quetico–Wawa, Winnipeg River – Wabigoon), evidence exists for earlier, intermediate-pressure (bathozone 5 and greater) metamorphism.

followed by voluminous plutonism in the intervals 2740–2695 and 2693–2680 Ma, again, generally showing a southward younging. Some rocks (Abitibi–Wawa, Quetico, Pontiac) had not yet formed when the 2705–2695 Ma activity began in the north. Thickening of the crust through thick-skinned deformation resulted in regional metamorphism and deformation (P_2–P_3). Melting of the lower crust, coincident with crustal thickening and promoted by thermal blanketing and several periods of crustal underplating, resulted in granulite-facies metamorphism (P_4–P_5) at both mid- and lower-crust levels and generated S-type granitic melts that were emplaced at upper- and mid-crust levels. Subsequent tectonism in the late Neoarchean and Paleoproterozoic served to uplift granulite-facies lower crust in various parts of the Superior Province and formed the present pattern of differentally exposed levels of the crust.

Economic implications

One aspect of the compilation project was to investigate relationships between metamorphism and mineralization, with the hope that new insights might be uncovered. Although complete investigation of this topic is beyond the scope of this paper, it is possible to draw some conclusions, and to highlight areas where further study would be fruitful.

Lode-gold deposits in the Superior Province commonly occur in greenschist- to lower-amphibolite facies rocks (*e.g.,* Fyon *et al.* 1992), locally in proximity to the greenschist–amphibolite boundary (*e.g.,* Red Lake, Andrews *et al.* 1986). Although the genetic significance of this spatial control of gold by metamorphic grade has not been firmly established, two main models have been proposed. 1) Prograde granulite-facies metamorphism in the lower crust results in the extraction of gold and associated lithophile elements (K, Rb), which are transported upward and precipitated into cooler crust (*e.g.,* Colvine *et al.* 1988, Cameron 1989, Corfu *et al.* 1989). 2) Gold is extracted locally from amphibolite-facies rocks, and is hydrothermally transported down a thermal gradient to sites of deposition (*e.g.,* Kerrich & Fryer 1979).

As has been noted previously (*e.g.,* Colvine *et al.* 1988, Corfu *et al.* 1989, Fyon *et al.* 1992), the timing of late Neoarchean gold mineralization within the Superior Province is generally consistent with this scenario. In fact, on the basis of the present data (*e.g.,* Fyon *et al.* 1992), there may have been two main pulses of gold mineralization, one event synorogenic, the other post-tectonic. Most attempts to date gold mineralization have focussed on the Abitibi greenstone belt, and commonly used Ar–Ar methods on micaceous minerals (*e.g.,* Masliwec *et al.* 1986, Hanes *et al.* 1992, Zweng *et al.* 1993, Powell *et al.* 1995). As noted previously herein, the post-tectonic metamorphic history of the Abitibi greenstone belt appears to exhibit a variety of localized thermal events (*e.g.,* Masliwec *et al.* 1986, Hanes *et al.* 1992, Zweng *et al.* 1993, Powell *et al.* 1995), making it difficult to ascertain whether or not gold mineralization occurred in a single, narrow time-interval (~10 Ma), or over a longer time-period (~40 Ma).

Gold mineralization may also have occurred prior to Neoarchean continental assembly (*e.g.,* Corfu & Stott 1991, Fyon *et al.* 1992). In the North Caribou greenstone belt, the presence of ~2850 Ma gold mineralization may be related to coeval high-grade metamorphism

TABLE 3. SUMMARY OF Ar–Ar HORNBLENDE AND BIOTITE AGES FROM THE SUPERIOR PROVINCE

Subprovince	Hornblende (in Ma)	Comment	Biotite (in Ma)	Comment	Cooling Rate
western Uchi	2675–2670	local preservation of M_1 cooling ages and ages of pluton crystallization	2660–2630	2660 in lower-greenschist rocks, 2630 Ma more typical regionally	2.5 to 3°C/Ma
central Abitibi	2670–2665	M_1, also local preservation of 2675–2705 Ma cooling ages and ages of pluton crystallization			2.5 to 3°C/Ma?
	2620	M_2, wide regional variation related to late hydrothermal and other events, local excess Ar	2500–2420	wide regional variation related to late hydrothermal and other events	2.5 to 3°C/Ma? + later local thermal effects
eastern English River – Winnipeg River	2660		2600		6 to 7°C/Ma?
western English River – Winnipeg River	2600		2470–2400		two-stage, pre- and post-2470 Ma
Wabigoon	2660–2630	few reliable ages	2575–2470	few reliable ages	2.5 to 3°C/Ma?
eastern Quetico	2600	few reliable ages	2570	few reliable ages	2.5 to 3°C/Ma?
south-western Abitibi (Levack gneiss)	2570		1800		history complicated by Sudbury event, multi-stage exhumation
Wawa gneiss domain	2570	excess Ar present	2480, 2300–2280	excess Ar present	slow cooling or multi-stage exhumation
Kapuskasing	2540, 2475–2500, ~2440	excess Ar common	2480, 2300–2280, 2230, 2000	excess Ar common	slow cooling or multi-stage exhumation.

Ages from Easton (1986), Feng *et al.* (1992), Hanes & Archibald (1999).

in the Wachusk metamorphic domain. Similarly, ~2740 Ma gold mineralization in the Pickle Lake camp (*e.g.,* Fyon *et al.* 1992) may be related to metamorphism of similar age in the southern Uchi subprovince. Thus, the mechanism of lode-gold mineralization at ~2850, ~2740 and ~2690–2650 Ma may all have been directly related to metamorphism in the lower crust.

Rare-element granitic pegmatite and fertile S-type granites in the Superior Province (*e.g.,* Fyon *et al.* 1992, Breaks & Tindle 1994, 1997, Tindle *et al.* 1998) may owe their development to the same processes that led to the formation of lode-gold deposits, namely, granulite-facies metamorphism in the lower crust resulting in the formation of granitic melts that were then transported upward into cooler crust along major structures, where they were emplaced, generally into amphibolite-facies rocks (Fyon *et al.* 1992). As with the deposition of gold, several pulses of pegmatite magmatism have been documented throughout the Superior Province between 2700 and 2640 Ma.

In the southern Superior Province, at least two main regional metamorphic events are recorded. The first is a higher-pressure event related to arc assembly and accretion; the second is a lower-pressure, late synplutonic event. As outlined by Powell *et al.* (1999) for the Hemlo area, and by Zaleski *et al.* (1999) for the Manitouwadge area, this complex tectonometamorphic history has served to mobilize and recrystallize early-formed ores. These studies provide a cautionary tale with respect to the application of relatively simple concepts of ore generation (*e.g.,* is the ore syngenetic or epigenetic) to greenstone belts with a polymetamorphic history.

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In a synthesis paper such as this one, it is difficult to acknowledge fully the large body of previous work which is drawn upon, particularly theses and detailed geological maps and reports by federal, provincial, and state surveys that form the backbone of our knowledge of the metamorphism of the province. References cited in the text do not necessarily represent a sole source of information, but also serve as an introduction to additional work cited therein. S. Buckley greatly assisted in the initial compilation of metamorphic data from the literature. K. Holland provided able assistance in digitizing the metamorphic polygons for the Superior and Southern provinces. Figures were drafted by E. Amyotte. T. Houlahan was responsible for the smooth exchange of digital data between the GSC and the OGS, and in translating the Ontario data to the map of the Canadian Shield. This paper has benefitted from comments by P. Thurston, and journal reviewers J. Percival and F. Corfu. This paper constitutes a product related to Ontario Geological Survey project PU94–07. It is published with permission of the Senior Manager, Precambrian Geoscience Section, Ontario Geological Survey.

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