

U–Pb systematics of garnet: dating the growth of garnet in the Late Archean Pikwitonei granulite domain at Cauchon and Natawahunan Lakes, Manitoba, Canada

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Abstract. This study considers the potential of using the U–Pb dating of garnet for determining quantitative P – T – t paths for the late Archean metamorphism in the Pikwitonei granulite domain. Garnets for U–Pb dating were selected mainly from samples that also provide information on pressure and temperature. The garnets used for dating were clear and free of any visible inclusions. Pb concentrations range from 63 ppb to 966 ppb and U from 136 ppb to 1143 ppb. The measured $^{206}\text{Pb}/^{204}\text{Pb}$ ratios range from 52.8 to 529.4. The ages are generally discordant with U/Pb ages that may lie above or below concordia. The discordance is caused by a recent disturbance of the U/Pb ratio in the garnets as indicated by replicate analyses on the same garnet separates that reproduce $^{207}\text{Pb}/^{206}\text{Pb}$ ages well within analytical uncertainty and in most cases within ± 1.5 Ma at 2600–2750 Ma. High grade metamorphism continued over a period of at least one hundred million years, but the garnet-K-feldspar Pb–Pb ages suggest that, during this time, garnet growth has been favored during three distinct periods in the Cauchon Lake area:

2700–2687 Ma

2660–2637 Ma

2605–2591 Ma

The ca. 2695 Ma garnet ages from Cauchon Lake date the time of melting and staurolite breakdown during prograde metamorphism, the ca. 2640 Ma ages date the time of extensive migmatization and the last period of metamorphic garnet growth, the ca. 2600 Ma ages date the time of crystallization of igneous garnet in late granitic intrusions. Peak metamorphism occurred around 2640 Ma followed by the intrusions of pegmatites starting at 2629 Ma. The Pb–Pb ages for garnet are similar to the U–Pb ages for zircon that date a leucocratic mobilizate (2695 Ma), a plagioclase-amphibole mobilizate (2637 Ma) and pegmatite (2598 Ma) (Heaman et al. 1986a; Krogh et al. 1986; this study). Xenocrysts of garnet from 2600 Ma old graphic granites give minimum ages of 2984 Ma and 2741 Ma which are minima for the times of garnet growth in the source of the granites. The agreement of the zircon and garnet ages suggests that the metamorphism may have been punctuated by events that led to the development of melts or encouraged mineral growth at specific times. If so, the prograde and retrograde paths of metamorphism in the area may have contained

minor excursions in pressure, temperature or fluid fugacities. In the Natawahunan Lake area some 50 km northwest of Cauchon Lake, garnet growth associated with the prograde breakdown of staurolite occurred at ca. 2744–2734 Ma. This suggests that a similar style of metamorphism may have occurred earlier in the Natawahunan Lake area than at Cauchon Lake area, or higher grades of metamorphism were reached earlier and were of longer duration associated with the somewhat greater depths in the Natawahunan Lake area. These results indicate that these garnets, which are 0.1–1 cm in diameter, have maintained closed system behavior for U and Pb at peak metamorphic conditions, i.e. temperatures up to 800° C and pressures of 7.5 kb.

Introduction

Several approaches have been developed that use mineral reactions and equilibria to determine pressure-temperature-(relative) time (P – T – t) paths for metamorphic terranes (Spear and Selverstone 1983; Droop and Bucher-Nurminen 1984; Waters 1986; Bohlen 1987). So far it has not been possible to precisely date the minerals involved in these reactions or equilibria because of either unfavorable daughter to parent ratios in the minerals or the relatively low temperatures of closure of the minerals considered.

In order to understand the evolution of a metamorphic terrane, it is essential to develop a method that will provide the precise time that a given mineral reaction or equilibrium occurred, especially when this information is used to help determine a P – T – t path. High precision U–Pb mineral ages have been obtained almost exclusively on accessory minerals, especially zircon, but also on sphene, monazite, xenotime, apatite and baddeleyite (Oosthuyzen and Burger 1973; Köppel 1974; Köppel and Grünenfelder 1975; Krogh et al. 1984; Corfu et al. 1985; Corfu and Grunsky 1987; Tucker et al. 1987).

Zircon has a relatively high temperature of closure and is thus very useful for dating. However, appropriate metamorphic reactions have not been described for zircon or most other accessory minerals, and it is presently not possible to determine when such accessory minerals formed along a given P – T – t path. This limits the usefulness of high precision U–Pb ages of accessory phases in a metamorphic terrane.

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The ideal mineral(s) for providing ages along $P-T-t$ paths should: be found in a range of common rock types, occur under a wide range of metamorphic conditions, have a high U/Pb ratio, have high closure temperatures for U and Pb gain or loss, and be one of the minerals involved in the reactions or equilibria used for determining P and T along pro- and retro-grade paths. In this study we demonstrate that garnet is such a mineral. Garnet is an important rock forming mineral in many metasedimentary and metaigneous rocks ranging from low to high grades of metamorphism and is used in many well-calibrated geothermometers and geobarometers (e.g. Essene 1982; Bohlen et al. 1983b). Most importantly, garnet is the prime mineral used in establishing $P-T-t$ paths (Hodges and Spear 1982; Bohlen 1987; Spear and Rumble 1987). It is also possible, based on petrography, to select samples that have garnets which formed by only one pro- or retro-grade reaction, thus reducing the potential for mixed ages.

U and Pb diffusion in garnet

While the diffusivity of U and Pb in garnets as a function of pressure, temperature and chemical environment is not known, field and limited experimental studies indicate that the rate of diffusion of cations in garnets is relatively low. In evaluating the diffusion data for garnet we will use the generalizations that for a given ionic radius, the element with the higher charge has the lower diffusivity, and for a given charge, the element with the larger ionic radius has the lower diffusivity (Freer 1981; Foland 1974).

Field and experimental studies show that garnets become homogeneous with respect to Fe and Mg above 650°–700° C (Yardley 1977; Cygan and Lasaga 1985). However, zoning of Ca in garnet may be preserved to much higher temperatures (Petraakis 1986). Diffusion data for Sm^{3+} at pressures of 30 kb (Harrison and Wood 1980) give closure temperatures of 440° to 570° C for pyrope and 670° to 760° C for grossular for a grain size of 0.4 cm and a cooling rate of 2°–100° C/Ma. The ionic radius of U^{4+} is similar to that of Ca^{2+} and Sm^{3+} (Shannon and Prewitt 1969). Therefore, U^{4+} with a higher charge should diffuse even more slowly and give higher closure temperatures than those for Ca^{2+} or Sm^{3+} .

Both garnets and zircons have a dodecahedral site of similar shape and size. In zircon this site strongly prefers U over Pb. U^{4+} should substitute for Ca^{2+} in the dodecahedral site of garnet and for Zr^{4+} in the dodecahedral site of zircon. For zircon there should be a greatly increased substitution of U^{4+} for Zr^{4+} , as compared to the substitution of U^{4+} for Ca^{2+} in garnet, because U and Zr have an identical charge and are similar in size. The incorporation of U^{4+} into the dodecahedral site of garnet may require the substitution of ions with a smaller charge in adjacent sites to maintain charge balance. During diffusion of U^{4+} such ions would have to diffuse as well. Minerals that show this type of coupled diffusion tend to preserve their original zoning to higher temperatures than equivalent minerals where ions of similar size and the same charge diffuse (e.g. alkali feldspars vs. plagioclase).

Because of its large ionic radius, the diffusion of Pb in zircon as well as in garnet is expected to be very slow. The major causes for Pb loss in zircons are considered to be volume diffusion (Tilton 1960) and alteration (Krogh and Davis 1974). Pb loss is pronounced in small, strongly

metamict grains which are susceptible to alteration. Clear zircons from Precambrian rocks with U concentrations of less than a hundred ppm have negligible metamictization and are generally retentive for U and Pb even at high grades of metamorphism (Percival and Krogh 1983). Pb loss can also be very small even in metamict zircons, if they are large or unaltered (see Table 1). Compared with most zircons used for dating, garnets are fairly large (several mm to a few centimeters). The concentrations of U in garnets are on the order of a few hundred ppb or less, therefore, metamictization and thus loss associated with alteration should be negligible. The main mechanism for loss of Pb from garnet should thus be volume diffusion, and the rate of loss should be low.

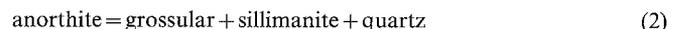
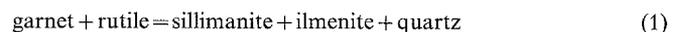
Geologic setting

Recognizing that garnets may provide useful U–Pb ages in metamorphic terranes, their utility for U–Pb dating was evaluated for Archean rocks from the Pikwitonei granulite domain and the Cross Lake subprovince along the northwestern margin of the Superior Province (Manitoba, Canada) (Fig. 1).

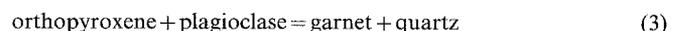
In the area of study, the tonalite gneiss-greenstone terrane grades east to west from amphibolite to granulite facies as a result of metamorphism that developed during the late Archean. The transition from amphibolite facies in the Cross Lake subprovince to granulite facies in the Pikwitonei domain is prograde and continuous and does not appear to be the result of structural or lithological breaks. The area is underlain dominantly by felsic gneisses, foliated to massive granitoids and amphibolites with subordinate amounts of arenaceous and argillaceous metasedimentary rocks, banded iron formation, metagabbros, calc-silicate rocks and marbles (Weber 1976; Weber and Scoates 1978; Weber and Loveridge 1981). Thus, the boundary between the Pikwitonei granulite domain and the Cross Lake subprovince in the Cauchon Lake area is based solely on the amphibolite to granulite transition, the orthopyroxene isograd, across a continuous sequence of rocks. At Cauchon Lake a greenstone belt strikes obliquely to this orthopyroxene isograd. Its eastern part has amphibolite facies assemblages, whereas the western equivalent has granulite facies assemblages (Fig. 1b).

During the Hudsonian orogeny the Churchill and Superior Provinces collided. During this orogeny the granulite facies rocks of the Superior Province were locally retrogressed to amphibolite and greenschist facies grade and partly migmatized within the Thompson mobile belt (Hubregtse 1980; Baragar and Scoates 1981). For this study no samples which showed any visible effect of lower grade Hudsonian overprint were analyzed.

The peak metamorphic temperatures (Fig. 1b and Table 5) are estimated from the two-feldspar geothermometer of Haselton et al. (1983). The temperatures are generally 750 to 800° C and ca. 640° C where retrogressed by the local intrusion of late granitic melts (Fig. 1b). The metamorphic pressures for metapelites (Fig. 1b and Table 5) are based on the following equilibria (Bohlen et al. 1983a; Edwards and Essene 1988):



Pressures for the mafic rocks were inferred using the equilibria (Bohlen et al. 1983b):



The activities of end-member garnet components were calculated with the model of Ganguly and Saxena (1984) with Fe–Ca interaction parameters from Anovitz and Essene (1987). The anorthite activity is based on the Al-avoidance model of Kerrick and Darken (1975) as formulated by Newton et al. (1980). The activity of ferrosilite in orthopyroxene was calculated according to the model of Wood and Banno (1973).

Table 1. Analytical results

| Sample | Pb (ppm) | U (ppm) | Atomic ratios | | | | | Ages in Ma | | | |
|-----------------|-------------|------------|---|---|---|--|--|--|--|---|-----------|
| | | | $\frac{^{206}\text{Pb}^a}{^{204}\text{Pb}}$ | $\frac{^{208}\text{Pb}^b}{^{206}\text{Pb}}$ | $\frac{^{207}\text{Pb}^b}{^{206}\text{Pb}}$ | $\frac{^{206}\text{Pb}}{^{238}\text{U}}$ | $\frac{^{207}\text{Pb}}{^{235}\text{U}}$ | $\frac{^{206}\text{Pb}}{^{238}\text{U}}$ | $\frac{^{207}\text{Pb}}{^{235}\text{U}}$ | $\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$ | |
| Garnets | | | | | | | | | | | |
| A-K | 225 | 0.121 | 0.529 | 151.7 | 0.1170 | 0.22044 (28) | 0.1520 (6) | 4.621 (20) | 912 | 1753 | 2984 ± 2 |
| A-L | 436 | 0.249 | – | 426.5 | 0.8066 | 0.18985 (18) | – | – | – | – | 2741 ± 2 |
| B-1 | 589 | 0.585 | 0.298 | 129.6 | 2.6224 | 0.19024 (27) | 0.5211 (11) | 13.668 (37) | 2704 | 2727 | 2744 ± 2 |
| B-2 | 589 | 0.498 | 0.352 | 143.0 | 2.9969 | 0.19020 (27) | 0.3497 (23) | 9.171 (61) | 1933 | 2355 | 2744 ± 2 |
| C-1 | 552 | 0.197 | 0.176 | 84.5 | 0.7303 | 0.18955 (54) | 0.4923 (28) | 12.866 (85) | 2580 | 2670 | 2738 ± 5 |
| C-2 | 552 | 0.606 | 1.143 | 71.39 | 1.8354 | 0.18910 (31) | 0.1499 (27) | 3.909 (71) | 900 | 1615 | 2734 ± 3 |
| D-1 | 476A | 0.213 | 0.262 | 108.6 | 0.2102 | 0.18523 (64) | 0.4723 (35) | 12.06 (10) | 2494 | 2609 | 2700 ± 6 |
| D-2 | 476A | 0.966 | 0.784 | 76.37 | 1.2640 | 0.17943 (30) | 0.4134 (30) | 10.226 (79) | 2230 | 2455 | 2648 ± 3 |
| E | 462 | 0.786 | 0.677 | 196.8 | 1.0616 | 0.18458 (22) | 0.5257 (19) | 13.380 (53) | 2723 | 2707 | 2694 ± 2 |
| F-1 | 217 | 0.654 | 0.528 | 456.4 | 1.4513 | 0.18394 (23) | 0.5227 (72) | 13.25 (18) | 2711 | 2698 | 2688 ± 2 |
| F-2 | 217 | 0.564 | 0.432 | 256.6 | 1.6052 | 0.18373 (14) | 0.5035 (15) | 12.755 (38) | 2629 | 2662 | 2687 ± 2 |
| F-3 | 217 | 0.734 | 0.603 | 529.4 | 1.5327 | 0.18372 (13) | 0.4983 (27) | 12.622 (68) | 2606 | 2652 | 2687 ± 1 |
| F-4 | 217 | 0.594 | 0.387 | 114.9 | 1.4726 | 0.18402 (62) | 0.5473 (21) | 13.887 (75) | 2814 | 2742 | 2689 ± 6 |
| G-1 | 157 | 0.137 | 0.174 | 86.10 | 2.5718 | 0.18078 (44) | 0.1995 (16) | 4.974 (41) | 1172 | 1815 | 2660 ± 5 |
| G-2 | 157 | 0.191 | – | 158.0 | 2.5320 | 0.1793 (18) | – | – | – | – | 2646 ± 17 |
| H-1 | 388 | 0.175 | – | 155.5 | 0.7070 | 0.18122 (20) | – | – | – | – | 2664 ± 2 |
| H-2 | 388 | 0.168 | 0.136 | 98.15 | 1.0714 | 0.17926 (40) | 0.4949 (16) | 12.233 (50) | 2591 | 2622 | 2646 ± 4 |
| H-3 | 388 | 0.063 | – | 82.61 | 0.3860 | 0.17896 (66) | – | – | – | – | 2643 ± 6 |
| I | 64 | 0.171 | 0.266 | 52.79 | 0.3246 | 0.17830 (51) | 0.2528 (75) | 6.21 (18) | 1453 | 2007 | 2637 ± 5 |
| K | 225 | 0.222 | – | 121.3 | 0.7560 | 0.17494 (43) | – | – | – | – | 2605 ± 4 |
| L | 436 | 0.101 | – | 313.7 | 0.6033 | 0.17343 (29) | – | – | – | – | 2591 ± 5 |
| M-1 | 254 | 0.187 | 0.271 | 284.2 | 1.4971 | 0.18493 (64) | 0.2792 (34) | 7.119 (92) | 1587 | 2126 | 2697 ± 6 |
| M-2 | 254 | 0.326 | 0.387 | 300.9 | 0.9873 | 0.18253 (21) | 0.4200 (36) | 10.572 (92) | 2261 | 2486 | 2676 ± 2 |
| M-3 | 254 | 0.481 | 0.311 | 156.8 | 1.9949 | 0.18069 (31) | 0.5027 (57) | 12.52 (14) | 2625 | 2644 | 2659 ± 3 |
| M-4 | 254 | 0.415 | 0.470 | 164.4 | 1.2914 | 0.17872 (49) | 0.3566 (52) | 8.78 (13) | 1966 | 2316 | 2641 ± 5 |
| Zircon | | | | | | | | | | | |
| N-1 | Z2 | 507 | 933 | 53265 | 0.0578 | 0.17745 (9) | 0.5110 (10) | 12.503 (25) | 2661 | 2643 | 2629 ± 1 |
| N-2 | Z2 | 1055 | 1952 | 30753 | 0.0505 | 0.17740 (13) | 0.5110 (21) | 12.499 (53) | 2661 | 2643 | 2629 ± 1 |
| O | Z1 | 1433 | 2799 | 40403 | 0.0264 | 0.17420 (9) | 0.4953 (14) | 11.897 (34) | 2594 | 2596 | 2598 ± 1 |
| Allanite | | | | | | | | | | | |
| P | Z1 | 1009 | 37.8 | 373.2 | 58.59 | 0.17440 (16) | 0.5140 (13) | 12.359 (33) | 2674 | 2632 | 2600 ± 2 |

^a Measured ratio

^b Corrected for mass fractionation, blank and common Pb

The pressures are ca. 7.0 kb at the transition from amphibolite to granulite facies at Cauchon Lake and ca. 7.5 kb for the highest grade granulites at Natawahunan Lake. Similar values were obtained from both pelitic and mafic gneisses using equilibria (1) and (3) (Fig. 1 b, Mezger et al., in preparation). Pressures estimated with equilibrium (2) (Table 5) show a larger scatter and are considered less reliable (Bohlen and Lindsley 1987).

Selection of garnets for dating

We have obtained ages for garnets from both upper-amphibolite and granulite facies rocks. Two samples, B, and C, are from metapelitic granulite grade rocks at Natawahunan Lake. Sample A and samples D through M are from metapelitic and metagneous amphibolite to granulite grade rocks as well as late granites in the Cauchon Lake area (Fig. 1, Table 1). In addition, U–Pb ages were

determined on zircons and one allanite from selected rock units within the amphibolite-granulite facies transition zone (Heaman et al. 1986a; Krogh et al. 1986; Table 1.). This permits a comparison of the U–Pb systematics for garnet and zircon.

Petrographic observation of ca. 500 samples revealed several reaction textures which can be used to establish the high-grade parts of the *P*–*T* paths for the Pikwitonei granulite domain. Based on the petrographic observations the *P*–*T* paths for the Cauchon Lake and Natawahunan Lake areas are very similar. These paths include an initial heating and burial stage during which the terrane passed from the andalusite into the sillimanite stability field, and reached temperatures above the stability of staurolite. This was followed by near-isobaric cooling (Mezger et al., in preparation). Most garnet samples used for dating were chosen to add absolute age information to these paths. Therefore samples were preferred where the garnet producing reaction can be inferred from textural

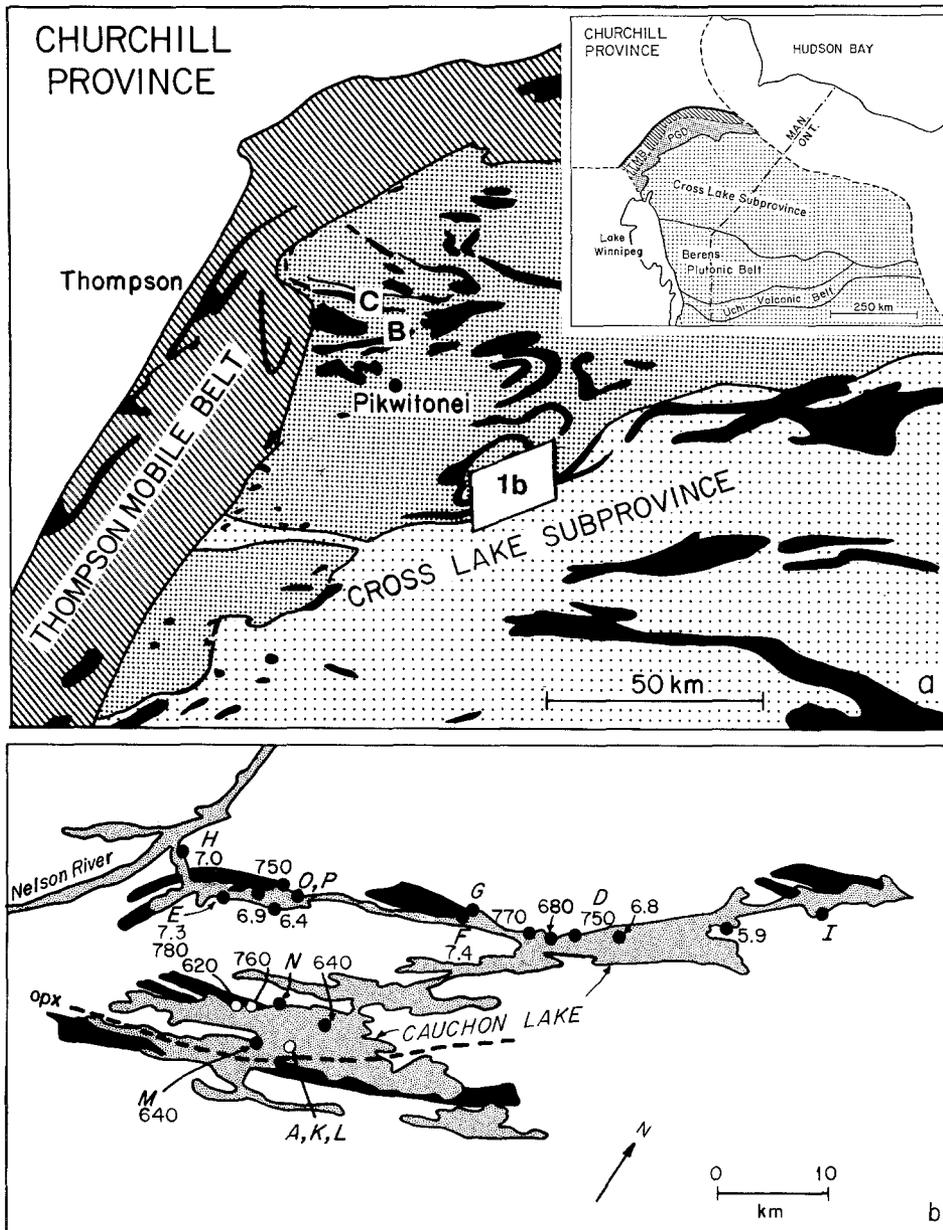


Fig. 1. **a** Geological map of the Pikwitonei granulite domain and the western parts of the Cross Lake subprovince (after Manitoba Mineral Resources 1980) with sample locations. Supracrustal units are shown in *black*. Most of the samples for this study are from Cauchon Lake (*box labeled 1b*, enlarged in Fig. 1b) along the border zone of the Cross Lake subprovince and Pikwitonei granulite domain. Two samples (*B* and *C*) are from the western part of the Pikwitonei granulite domain at Natawahunan Lake. **b** This map of the Cauchon Lake area (after Weber 1976; Weber and Loveridge 1981) shows the locations for the samples used for dating (*capital letters*). Also shown are the temperatures (in degrees) obtained using the two-feldspar thermometer and the pressures (in kb) estimated from the equilibria GRAIL and garnet-orthopyroxene-plagioclase-quartz. The *dashed line* represents the "orthopyroxene-in" isograd after Weber and Loveridge (1981), which separates the Cross Lake subprovince to the south-east from the Pikwitonei granulite domain. *Black areas* indicate supracrustal units

criteria. The mineralogy and exact location of individual samples selected for analyses are given in the Appendix.

An individual garnet may form by different reactions and over an extended period of time. If the closure temperature for U and Pb loss or gain is higher than the temperature achieved during metamorphism, the ages determined would then be representative of the total period during which the garnet grew. Within a single hand specimen several generations of garnets may be distinguished petrographically and each of these generations may have developed at different times. Because different generations of garnet may exist within an individual sample, single garnets were selected for analysis whenever possible. If more material was needed, garnets were taken only from a single lithological layer if the sample was heterogeneous, as it was the case for most metapelites. Garnets were taken from crushed whole rocks if the samples were homogeneous and only one type of garnet was observed to be present.

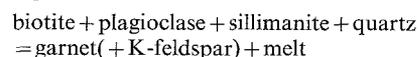
In evaluating whether garnet could be used for dating, the first garnets analyzed, M-2 and H-1, were selected from crushed whole rock samples that exhibited a pronounced compositional layering. As a result, the ages obtained for these two samples could

be mixed ages. For the subsequent samples, the garnets selected generally fulfilled one of the following criteria:

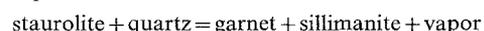
1. Garnet formed by a mineral reaction that can be observed in thin section. (samples B, C, F)
2. Garnet formed by a reaction involving partial melting or crystallized from a granitic melt. (samples D, K, L, M-4)
3. Only one type of garnet occurs in a rock with a homogeneous mineralogy or within an individual layer of a heterogeneous sample. In such cases it is likely that all garnet formed by the same reaction, but the reactants can not be observed petrographically. (samples E, G, H, I).

Textural criteria indicate that some of the garnets analyzed formed by the following melting and dehydration reactions:

Separates D, M-4:



Separates B, F:



Separate C:

staurolite = garnet + sillimanite + spinel + vapor

Analytical techniques

The rocks or individual garnets were crushed to 0.85–0.18 mm and purified with heavy liquids and/or a Frantz Isodynamic magnetic separator. Garnet fragments without visible inclusions were handpicked under a binocular microscope. Common inclusions included biotite, feldspar, spinel, sillimanite, graphite and rarely zircon, rutile and monazite. To evaluate whether inclusions not visible with an optical microscope were present, some garnets were inspected with the backscattered electron imaging on the electron microprobe. No inclusions were detected in the handpicked separates used for dating.

Some 20 to 150 mg of fresh garnet fragments were picked for each analysis and washed in warm, doubly distilled 2 N HCl for about 20 min, to remove surface contamination. Samples crushed to ca. 0.075 mm and washed in hot 2N HCl for 1 h are among the most discordant (samples A–K and I) suggesting that Pb can be leached by hot weak acid, if the grain size is small.

The samples were digested in screw-top, Teflon PFA (Savilex®) vials on a hot-plate with an HF-HNO₃ mixture and a few drops of concentrated H₂SO₄. Decomposition of the garnets into a white precipitate, most likely fluorides, was achieved at ca. 150° C within 6 to 24 h depending on grain size. After evaporating to dryness for 10–15 h at ca. 200° C, the residue was completely soluble in hot 6N HCl. The sample remained in solution when diluted to 2N HCl for ion exchange chemistry. Because of their high viscosity at room-temperature, the solutions were loaded hot on the ion-exchange columns. The high viscosity indicates that the solutions are not purely ionic, but may contain a significant gel-component. In the early work, the sample was aliquoted after dissolution and one part was spiked with a ²⁰⁸Pb–²³⁵U mixed spike. When a ²⁰⁵Pb–²³⁵U spike became available (provided by T. Krogh and R. Parrish, Royal Ontario Museum, Toronto, Canada), the spike was added before HF digestion. Whole rock samples were decomposed in a HF–H₂SO₄ mixture in Krogh-style bombs. Pb was separated by HCl–HBr chemistry and U by HCl–HNO₃ chemistry using Dowex 1–X8 resin (Tilton 1973; Manhès et al. 1984; Mattinson 1986). The spikes were calibrated against two different standard solutions, one was prepared at Stony Brook, the other one was provided by K. Ludwig, USGS, Denver, USA. The U/Pb ratios in the spikes for both calibrations agree to within ±0.15%. The reproducibility of the U/Pb ratio with a single standard solution is ±0.05%.

Leached K-feldspar from each sample was analyzed for its Pb isotope composition, which then was used for the common Pb correction for the coexisting garnet. K-feldspar was crushed to 0.36–0.18 mm and separated with heavy liquids and a Frantz Isodynamic magnetic separator. 5–30 mg of fresh, inclusion-free fragments were handpicked under the binocular microscope. It has been shown that leaching K-feldspars removes their more radiogenic Pb-component (Ludwig and Silver 1977). This radiogenic Pb can be removed because it is probably located in fractures (in silicate alteration phases such as micas) or in U–Th rich inclusions (such as uraninite, uranothorite, thorianite, etc.) in K-feldspars (J. Stuckless, personal communication; Krogstad et al., in preparation). Therefore all K-feldspars were washed in boiling 6N HNO₃ and in 6N HCl for at least 30 min to remove any surface contamination and possible radiogenic Pb in cracks or altered zones. Subsequently the K-feldspars were leached with HF of increasing strength and some HNO₃. After two to three leaching steps the residue was dissolved in concentrated HF only. HNO₃ was added to the leaching steps to dissolve any potential U–Th bearing phases in the K-feldspar grains that were invisible with the binocular microscope and were not removed during the washing steps. HNO₃ was not used for the final dissolution step to prevent the complete dissolution of these phases. Pb was separated by HCl–HBr chemistry. For all samples in Table 1, the Pb isotopic ratios for the garnets were corrected for inherited Pb using the

Pb ratios obtained from the K-feldspar residue (Table 2). For zircon analyses, Pb and U were separated using the technique of Krogh (1973).

For isotope measurements Pb was loaded on a rhenium center filament using the phosphoric acid-silica gel method (Cameron et al. 1967). U was loaded with phosphoric acid on a rhenium side filament using a rhenium center filament for ionization, or with phosphoric acid and graphite on a single rhenium filament. The total procedural blank for Pb was 300 to 110 pg for garnets and 100 pg for K-feldspars and zircons. A typical garnet sample consist of 15 to 50 ng of Pb, thus the blank is usually less than 1% of the sample. No blank correction was necessary for the K-feldspars. The measured Pb ratios for the blank were:

$$^{206}\text{Pb}/^{204}\text{Pb} = 18.05 \pm 0.5$$

$$^{207}\text{Pb}/^{204}\text{Pb} = 15.66 \pm 0.1$$

$$^{208}\text{Pb}/^{204}\text{Pb} = 37.86 \pm 0.5$$

All isotope ratios were measured with a Faraday collector. Two nanograms of Pb give a signal intensity of 1×10^{-11} A. Most Pb ratios measurements were made with a signal of 1 to 5×10^{-11} A for the largest peak. The mass discrimination factor for Pb is $0.10 \pm 0.02\%$ per AMU and the uncertainty (two standard deviations) for any one analysis of the ²⁰⁷Pb/²⁰⁶Pb ratio is 0.05% based on repeated analyses of NBS equal atom standard SRM 982. The uncertainty in the ²⁰⁶Pb/²⁰⁴Pb ratio is 0.1% for ratios less than 100.

The total procedural blank for U was 50 pg. No mass discrimination factor was applied to the U isotopic ratios, based on repeated analyses of NBS-SRM U-930. Decay constants are those of Jaffey et al. (1971) as recommended by the International Union of Geological Sciences (IUGS) (Steiger and Jäger 1977).

The measured concentrations of Pb and U range from 63 to 966 ppb and 136 to 1143 ppb, respectively. The measured ²⁰⁶Pb/²⁰⁴Pb ratios range from 52.9 to 530.5 (Table 1). These ratios are similar to ratios obtained on some sphenes (Corfu and Grunsky 1987; Tucker et al. 1987) and apatite (Oosthuyzen and Burger 1973), but are lower than those generally reported for zircons or monazites.

The uncertainties (2σ) for the U/Pb and ²⁰⁷Pb/²⁰⁶Pb ratios reported in Table 1 were calculated according to the method of Ludwig (1980, 1982). The uncertainties include the reproducibility of the standard ratios, the effect of fractionation correction, blank, common Pb correction based on the composition of the coexisting K-feldspar and uncertainties due to within-run statistics. The uncertainties for the ²⁰⁷Pb/²⁰⁶Pb ages are calculated directly from the uncertainty in the reported ²⁰⁷Pb/²⁰⁶Pb ratios. The uncertainty in the ²⁰⁷Pb/²⁰⁶Pb age based only on the uncertainties in the measured isotope ratios in the garnet and K-feldspar is mainly dependent on the ²⁰⁶Pb/²⁰⁴Pb ratio of the garnet. For a 2700 Ma old garnet, a ²⁰⁶Pb/²⁰⁴Pb of 50 results in an uncertainty in the ²⁰⁷Pb/²⁰⁶Pb age of 4 Ma, at a ratio of 100 an uncertainty of 2 Ma and at ratios of about 500 or greater the uncertainty depends mainly on the uncertainty in the measurement of the ²⁰⁷Pb/²⁰⁶Pb ratio of the garnet and is about 1 Ma. The reported uncertainties are usually somewhat large due to the uncertainty in the blank correction or, for small samples, unfavorable within-run statistics.

The garnets were analyzed for major elements using a CA-MECA-CAMEBAX microanalyser employing wavelength dispersive spectrometers equipped with LiF, PET and TAP crystals. Analyses were performed with an acceleration voltage of 15 kV and a beam current of 25 nA. Spectrometer data were corrected for atomic number, absorption, fluorescence and backgrounds using the CAMECA ZAF correction program.

Common lead correction

For samples with low ²⁰⁶Pb/²⁰⁴Pb ratios (Fig. 2), the effect of the corrections for inherited Pb upon the calculated ages is significant, and it is, therefore critical to know the appropriate isotope ratios for the inherited Pb component. These ratios are best estimated from analyses of minerals with

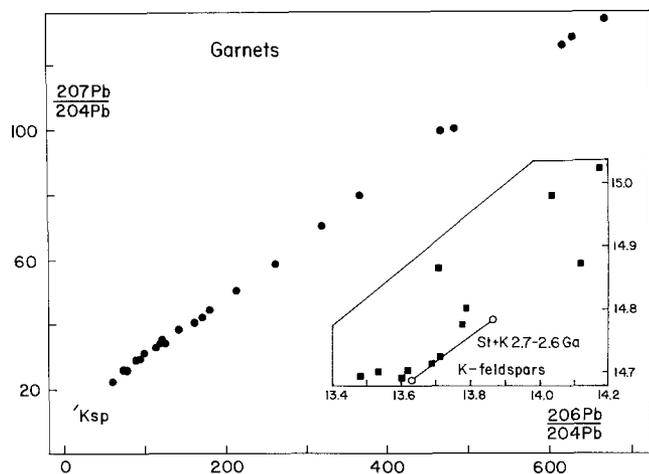


Fig. 2. $^{206}\text{Pb}/^{204}\text{Pb}$ vs. $^{207}\text{Pb}/^{204}\text{Pb}$ diagram showing the blank- and fractionation-corrected Pb isotope ratios for the garnets analyzed (closed circles). Also shown is the total variation of the Pb isotopic ratios of leached K-feldspars (Ksp) coexisting with the garnets. The inset shows the Pb ratios for individual analyses of leached K-feldspars (closed squares) and the Stacey and Kramers (1975) Pb-evolution model from 2.7 to 2.6 Ga (line connecting open circles labelled St + K 2.7-2.6 Ga)

low U/Pb ratios separated from the rock or using the Pb isotope ratios of the whole rock. For calculating the U–Pb ages, both the Pb isotopic composition and the U/Pb ratio of the whole rock or the mineral used for inherited Pb correction are required. We decided to use the Pb isotope composition of leached K-feldspars for the common Pb correction (Table 2, Fig. 2). K-feldspars preferentially incorporate Pb over U. Leaching K-feldspar with HF has been shown to remove its radiogenic Pb component, so that the residue should give a good approximation of the Pb isotope composition for the K-feldspar at the time of its closure (Ludwig and Silver 1977). However, based on experimental diffusion data (Rosenqvist 1949), Pb diffuses more easily in K-feldspar than in garnet. Therefore, the Pb-isotopic composition of K-feldspar should be more readily reset than the Pb-isotopic composition of garnet. The isotopic ratios measured for K-feldspar may reflect the ratios of the whole rock during the last thermal disturbance and may therefore differ from those at the time of garnet growth. Whole rock-leached K-feldspar Pb–Pb ages for the three samples shown in Table 2 range from 3100 Ma to 2800 Ma with uncertainties of one to three hundred million years. These ages indicate that the whole rock–K-feldspar system did not get reset by post-Archean events such as the intrusion of the mafic Molson Dyke swarm at 1883 ± 2 Ma (Heaman et al. 1986b) or the collision of the Superior Province with the Churchill Province at ca. 1700 Ma.

A K–Ar amphibole age of 2430 Ma (Burwash et al. 1962) and rutile U–Pb ages of 2360 Ma from Natawahunan Lake and 2430 Ma from Cauchon Lake (K. Mezger, in preparation) probably date the time when the terrane cooled below ca. 500°C. The slow cooling rate of the terrane following granulite grade metamorphism, as indicated by these ages, did not cause a significant re-equilibration between the whole rocks and K-feldspars upon cooling. The K-feldspars are dominantly preserved as perthitic orthoclase and are only very rarely retrogressed to microcline.

Table 2. Pb isotope composition of whole rocks (WR) and leached K-feldspars (K-spar)

| Sample | Material analyzed | $^{206}\text{Pb}/^{204}\text{Pb}$ | $^{207}\text{Pb}/^{204}\text{Pb}$ | $^{208}\text{Pb}/^{204}\text{Pb}$ |
|------------------|-------------------|-----------------------------------|-----------------------------------|-----------------------------------|
| Natawahunan Lake | | | | |
| B | K-spar | 14.179 | 15.026 | 33.817 |
| C | K-spar | 13.486 | 14.695 | 33.068 |
| Cauchon Lake | | | | |
| A-K, K | K-spar | 13.716 | 14.727 | 33.377 |
| A-L, L | K-spar | 13.685 | 14.735 | 33.363 |
| D | K-spar | 14.044 | 14.980 | 33.923 |
| E | K-spar | 13.605 | 14.693 | 33.317 |
| E | WR | 14.580 | 14.920 | 35.043 |
| F | K-spar | 13.622 | 14.706 | 33.416 |
| F | WR | 14.865 | 14.959 | 35.388 |
| G | K-spar | 14.123 | 14.872 | 33.989 |
| H | K-spar | 13.538 | 14.703 | 33.322 |
| I | K-spar | 13.715 | 14.867 | 33.093 |
| I | WR | 14.118 | 14.947 | 34.386 |
| M | K-spar | 13.780 | 14.778 | 33.451 |
| O, P | K-spar | 13.792 | 14.804 | 33.584 |

This lack of isotopic and structural re-equilibration of the K-feldspars, despite the slow cooling rate, may be due to the lack of post-metamorphic deformation and the absence of a fluid phase, as indicated by the absence of hydrous retrogression in the rocks.

Based on the Pb isotope compositions of the whole rocks and K-feldspars, the rocks had $^{238}\text{U}/^{204}\text{Pb}$ (μ) ratios of less than 2.5 since the Archean and their Pb isotope ratios are therefore not much different from those of the leached K-feldspars. These low μ -values suggest that the terrane underwent U loss during the Archean. U loss is generally associated with granulite grade metamorphism. As discussed later in this paper, the thermal event around 2600 Ma reached only amphibolite grade and caused only localized retrogression of earlier granulite facies assemblages by the emplacement of granitic melts. Thus the latest time for granulite grade metamorphism is ca. 2640 Ma. This implies that the rocks must have attained their low μ -values before or during the granulite facies metamorphism at ca. 2640 Ma. The change in Pb isotopes for the whole rock from 2640 Ma to 2600 Ma is negligible. Assuming that the K-feldspar Pb-ratios were reset at 2640 Ma, and that the whole rocks had $^{238}\text{U}/^{204}\text{Pb}$ ratios of 8 from 2984 Ma, 2740 Ma or 2690 Ma to 2640 Ma and $^{238}\text{U}/^{204}\text{Pb}$ ratios ≤ 2.5 from 2640 Ma to 2600 Ma, there is no significant effect (i.e. < 1 Ma) on the garnet ages except for samples A–K, B and C, where the K-feldspar–garnet ages are increased by 2 to 10 Ma. For the samples A–K, B and C the ages reported in Table 1 are minimum ages and would have to be increased by an amount appropriate for the Pb-evolution in these rocks. Correction for the common lead component in sample C with a less evolved isotopic composition would increase the agreement in the ages of the duplicate analyses. The agreement of the K-feldspar–garnet ages within 1.4 Ma for the four replicates from sample F, despite the large variability in the $^{206}\text{Pb}/^{204}\text{Pb}$ ratios, indicates that the common Pb correction used for the garnets is valid.

Use of a model-mantle Pb, such as the Stacey and Kramers (1975) model, to correct for initial Pb in these garnets is not valid because the Pb for the various samples appears to have evolved to different extents in crustal environments (Fig. 2). For example, sample D has a leached K-feldspar with one of the more radiogenic $^{206}\text{Pb}/^{204}\text{Pb}$ ratios and a garnet with one of the lower $^{206}\text{Pb}/^{204}\text{Pb}$ ratios. Using the Stacey and Kramers model Pb for correction yields an age 17 Ma older than the age obtained using the Pb ratios from co-existing K-feldspar. However, for sample F-4, which has a $^{206}\text{Pb}/^{204}\text{Pb}$ ratio similar to that of sample D, the difference in age is only 0.5 Ma, because the Pb from the leached K-feldspar plots close to the Stacey and Kramers model curve. Thus, to obtain geologically meaningful and precise ages, it is critical to estimate for each sample, as precisely as possible, the isotopic ratios to be used for the inherited Pb component.

Location of U and Pb in garnet

The significance of the ages deduced from garnet may be dependent on the location of U and Pb in the garnet. Even though inclusions of zircon, rutile and rarely monazite occur in garnets from this area, several lines of evidence suggest that the U and Pb extracted from the handpicked garnets used for dating come from the garnet crystal structure rather than from inclusions.

As noted above, the dodecahedral site in garnet is almost identical in size and shape to the Zr site in zircon (Novak and Gibbs 1971). This suggests that U may be accommodated by the garnet structure. The abundances of U in the garnets are a few hundred ppb, much lower than for REE of similar ionic radii. The $^{208}\text{Pb}/^{206}\text{Pb}$ ratios for the garnets, corrected for common lead and blank, are 0.12 to 3.0. These $^{208}\text{Pb}/^{206}\text{Pb}$ ratios are higher than those

commonly observed for zircon and primary rutile and lower than those commonly observed for monazite. Samples B and G have $^{208}\text{Pb}/^{206}\text{Pb}$ ratios which are at the low end for ratios measured for monazite. Garnets from these two samples were studied using backscattered electron imaging and no evidence of monazite was found.

Despite our best efforts to find small inclusions, it is possible that submicroscopic inclusions could still exist somewhere in the garnet fragments used for dating. If so, it is most likely that submicroscopic grains would have formed at the same time as the garnet. Once included in garnet, they would be isolated. If U and Pb were mobilized, they would have to diffuse through the garnet and thus would yield ages that are the same as if the U and Pb were incorporated in the garnet. That similar amounts of inclusions would be incorporated in each garnet separate to yield similar U–Pb levels seems very unlikely. This, coupled with Pb-isotopic ratios inappropriate for zircon, rutile and in most cases monazite and the crystal chemical constraints leads us to conclude that the U and Pb extracted from the garnets were part of the garnet crystal structure.

The garnets show significant variability in U/Pb ratios and therefore in their $^{206}\text{Pb}/^{204}\text{Pb}$ ratios (Fig. 2). This variability does not correlate with the major element composition of these almandine-pyrope garnets (Table 3). All garnets are homogeneous with respect to Fe and Mg, except for a rim, a few tens of microns thick, that is enriched in Fe. Such a rim was included in most of the analyzed samples. However, because it is so thin compared to the 0.1 to 1 cm dimensions of the garnet crystals, it would have made up much less than 1% of any sample. Strongly peraluminous rocks and pegmatites were found to have garnets with the highest U/Pb ratios, probably reflecting the higher U/Pb ratios for these rock-types. The modal abundance of garnet in a rock does not appear to control the concentration of U and Pb or the U/Pb ratio in the garnets.

Table 3. Garnet major element analyses

| Sample | B | C | D | E | F | G | H | I | L | M-1 | M-4 |
|--------------------------------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|
| SiO ₂ | 38.90 | 38.79 | 39.02 | 38.83 | 39.09 | 39.22 | 39.36 | 37.96 | 37.83 | 37.65 | 37.65 |
| Al ₂ O ₃ | 22.49 | 21.88 | 21.71 | 22.28 | 21.79 | 22.05 | 22.05 | 21.26 | 20.70 | 20.96 | 21.20 |
| FeO | 27.09 | 27.64 | 28.57 | 27.65 | 27.85 | 31.36 | 30.25 | 32.60 | 33.58 | 34.89 | 34.25 |
| MgO | 9.96 | 8.95 | 9.50 | 9.51 | 8.49 | 7.49 | 8.46 | 5.76 | 4.10 | 4.52 | 5.07 |
| CaO | 1.33 | 1.10 | 0.95 | 1.02 | 1.12 | 1.84 | 1.40 | 1.53 | 1.60 | 0.56 | 0.50 |
| MnO | 0.45 | 0.58 | 0.62 | 0.97 | 0.74 | 0.62 | 0.63 | 1.60 | 2.87 | 1.80 | 1.74 |
| Total | 100.22 | 98.94 | 100.37 | 100.26 | 99.08 | 102.58 | 102.15 | 100.71 | 100.68 | 100.38 | 100.41 |
| Cations based on 24 oxygens | | | | | | | | | | | |
| Si | 5.958 | 6.028 | 6.000 | 5.964 | 6.072 | 5.983 | 5.992 | 5.978 | 6.029 | 6.007 | 5.984 |
| Al | 4.054 | 4.007 | 3.935 | 4.032 | 3.989 | 3.964 | 3.955 | 3.946 | 3.886 | 3.942 | 3.972 |
| Fe ²⁺ | 3.465 | 3.592 | 3.674 | 3.551 | 3.618 | 3.988 | 3.851 | 4.293 | 4.474 | 4.656 | 4.552 |
| Mg | 2.271 | 2.072 | 2.178 | 2.176 | 1.965 | 1.704 | 1.920 | 1.352 | 0.974 | 1.076 | 1.200 |
| Ca | 0.218 | 0.183 | 0.157 | 0.167 | 0.186 | 0.300 | 0.229 | 0.259 | 0.273 | 0.096 | 0.085 |
| Mn | 0.058 | 0.077 | 0.081 | 0.126 | 0.097 | 0.080 | 0.082 | 0.213 | 0.388 | 0.243 | 0.235 |
| Total | 16.024 | 15.959 | 16.025 | 16.016 | 15.927 | 16.019 | 16.029 | 16.041 | 16.024 | 16.020 | 16.028 |
| Endmember garnets | | | | | | | | | | | |
| Alm | 0.577 | 0.606 | 0.603 | 0.590 | 0.617 | 0.657 | 0.633 | 0.702 | 0.733 | 0.767 | 0.750 |
| Pyrop | 0.378 | 0.350 | 0.358 | 0.362 | 0.335 | 0.281 | 0.316 | 0.221 | 0.159 | 0.177 | 0.198 |
| Spess | 0.010 | 0.013 | 0.013 | 0.021 | 0.017 | 0.013 | 0.013 | 0.035 | 0.064 | 0.040 | 0.039 |
| Gross | 0.036 | 0.030 | 0.025 | 0.028 | 0.032 | 0.049 | 0.038 | 0.042 | 0.044 | 0.016 | 0.014 |

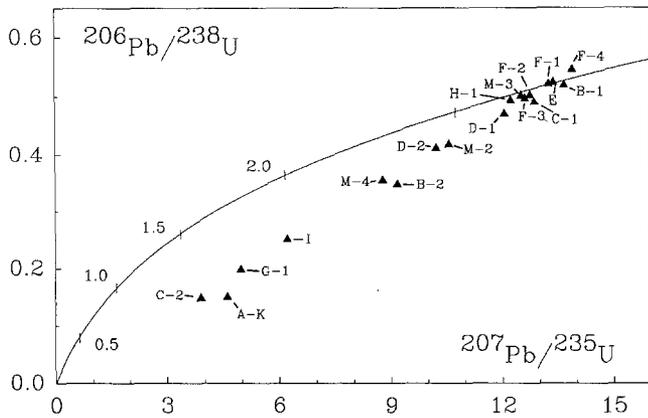


Fig. 3. U-Pb concordia diagram with corrected U-Pb data for garnet showing the extent of discordance. The dimensions of the symbols exceeds in most cases the analytical uncertainties

Results and discussion

Evaluation of U-Pb ages for garnet

As shown in Fig. 3 and Table 1 some of the garnets give concordant U-Pb ages, most of the ages are discordant, however. Most samples show normal, some show reverse, discordance. To test the reasons for the discordance, garnets for most samples were analyzed in replicate. The replicates represent two or more batches of the same mineral separate (samples B, C, F, G, H, except batch H-1), or different separates and different individual garnets from the same samples. The mineral separates represent garnets from whole rocks with a homogeneous mineralogy (samples B, F and G), individual layers of a heterogeneous rock (samples C, H and M-4) or individual garnets (A-K, A-L, K, L, D, M-1, M-3). For samples G and H the sample size was insufficient for a replicate U analysis in some cases. The $^{207}\text{Pb}/^{206}\text{Pb}$ ages agree within the analytical uncertainties only for samples where the same batch could be used for all the analyses, and only these analyses can be considered to be real replicates. In most of these cases the $^{207}\text{Pb}/^{206}\text{Pb}$ ages agree within ± 1.5 Ma. For samples where a different separate, i.e. a different individual garnet or a batch from a different part of a heterogeneous rock, had to be used the $^{207}\text{Pb}/^{206}\text{Pb}$ ages are usually significantly different.

The replicate analyses show that the discordance within a single sample can be quite variable despite the agreement of the $^{207}\text{Pb}/^{206}\text{Pb}$ ratios. The reproducibility of the $^{207}\text{Pb}/^{206}\text{Pb}$ garnet ages for the same garnet separates, despite the variable discordance indicates that the discordance is recent. This implies that the $^{207}\text{Pb}/^{206}\text{Pb}$ ages can be interpreted as the real ages for the garnets.

The discordance could be a result of weathering or the analytical procedure. All rocks were taken from surface exposures, but garnet samples showing any kind of weathering were not used for analyses. Also during handpicking, fragments with freshly broken surfaces were preferred. We consider it therefore more likely that the U/Pb ratios were changed during analysis. After dissolution of the garnets the solution have high viscosity probably due to the formation of a gel. This gel may make it difficult to ensure complete homogeneity of U and Pb between sample and spike.

Assuming that the garnet U/Pb ages are discordant due to recent changes in the measured U/Pb ratios, the $^{207}\text{Pb}/$

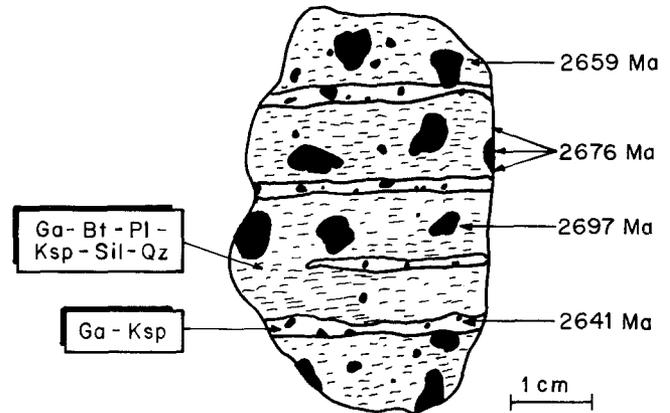


Fig. 4. Sketch of migmatite sample M. Four different garnet separates were dated from this sample. Small garnets exclusively separated from the leucosome give a Pb-Pb age of 2641 Ma. An individual garnet from the border leucosome-melanosome yielded an age of 2659 Ma. An age of 2676 Ma was obtained for a garnet separate from the whole rock. Corroded garnets from the melanosome which were completely enclosed in biotite, yielded an age of 2697 Ma

^{206}Pb age should give the times of garnet growth or the times for closure of U and Pb diffusion in the garnets.

As shown in Table 1 and Table 4 the $^{207}\text{Pb}/^{206}\text{Pb}$ ages cluster in the following time periods:

| Cauchon Lake | Natawahunan Lake |
|----------------|------------------|
| ≥ 2984 Ma | |
| ≥ 2741 Ma | |
| 2700-2687 Ma | 2744-2734 Ma |
| 2660-2637 Ma | |
| 2605-2591 Ma | |

The oldest and youngest ages from Cauchon Lake were obtained from garnets extracted from two hand specimens of late, coarse grained and partly graphic granites which intruded at the orthopyroxene isograd after granulite facies metamorphism (Fig. 1b). The minimum $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2984 Ma for sample A-K was obtained from an individual garnet, ca. 1 cm in diameter from sample K. Smaller garnets (≤ 0.3 cm in diameter) from the same hand specimen yield a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2605 Ma. The $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2741 Ma for sample A-L was obtained from a garnet extracted from specimen L. The same sample also yielded garnets with an age of 2591 Ma.

Two euhedral garnets from the leucosome of a nebulitic migmatite give ages of 2700 Ma for D-1 and 2647 Ma for D-2. In migmatite sample M, three different garnets can be distinguished petrographically (Fig. 4): corroded garnets, a large round variety (0.5 to 0.8 cm) and small grains (< 0.2 cm) with well developed crystal faces. The corroded garnets from the melanosome are completely enclosed by biotite and, based on textural criteria, appear to be the oldest generation of garnet within the sample. The melanosome contains all three types of garnet. The leucosome contains only small garnets with well developed crystal faces, although the large round garnets may grow from the melanosome into the leucosome. Based on petrographic observation, the garnets in the leucosome formed by a reaction producing garnet + K-felspar \pm melt. Four different garnet separates were picked from the sample and they give four distinct ages (Fig. 4):

1. Corroded garnets for the malanosome: 2697 Ma
2. Small garnets with well developed crystal faces from the leucosome: 2641 Ma
3. A mixture of garnets separated from the whole rock: 2676 Ma
4. A single large, round garnet from the border of the leucosome with the melanosome: 2659 Ma.

The occurrence of garnets with multiple ages in the late graphic granites and migmatites, where their relative ages can be predicted based on petrographic observation, suggests that U–Pb ages for garnet are not reset by later partial melting. The garnets in sample F formed by the prograde break-down of staurolite + quartz at ca. 700° C (Richardson 1968). Thus granulite grade conditions must have been reached at Cauchon Lake after the formation of these garnets and before the beginning of intrusion of late pegmatites at 2629 Ma (Table 1). Since some garnets from the late granites record ages which are much older than those for sample F, their $^{207}\text{Pb}/^{206}\text{Pb}$ ages must have also survived granulite-grade metamorphism. During granulite facies metamorphism temperatures of 750°–800° C were reached, as recorded by the two-feldspar geothermometer. Therefore the closure temperature for U/Pb diffusion in garnet has to be higher than 750°–800° C. These high closure temperatures indicate that the $^{207}\text{Pb}/^{206}\text{Pb}$ ages are the growth ages of the garnets. This makes garnets particularly useful for understanding the prograde evolution of a metamorphic terrane (Table 4).

Therefore, the 2744 Ma to 2734 Ma ages for the garnets from Natawahunan Lake are interpreted to date the time of garnet growth at relatively high temperature, i.e. above the stability of staurolite + quartz and staurolite. These ages are significantly older than the 2687 Ma age for garnets from sample F, that developed by the break-down of staurolite + quartz in the Cauchon Lake area. However, an age of 2741 Ma is found for a garnet xenocryst from a 2600 Ma graphic granite from the Cauchon Lake area.

The minimum ages of 2984 Ma and 2741 Ma from granite samples K and L are interpreted to place a minimum on the times of garnet growth in the source of the granites. The younger ages of 2605 Ma and 2591 Ma are interpreted to represent the time of intrusion of these coarse-grained granites.

A possible interpretation for the garnets in the leucosome of sample D is that the rock underwent partial melting around 2700 Ma and again at ca. 2648 Ma when parts of the older leucosome were remobilized during granulite grade metamorphism. The garnet ages of 2700 Ma and 2648 Ma from this sample may have to be interpreted as minimum or maximum ages, respectively, depending on the amount of new garnet overgrowth or inheritance.

Garnets grew at 2660 Ma in a metamorphic pegmatite (sample G), at 2646 in parts of metapelite H and at 2637 Ma in metarhyolite I. The reactions by which the garnets from these three samples formed cannot be reconstructed from petrographic observations.

For migmatite sample M (Fig. 4) we interpret the age of 2697 Ma to represent an early metamorphism when the first generation of garnets in the melanosome formed. During granulite grade metamorphism at 2641 Ma the garnet-bearing leucosome developed. The age of 2676 Ma determined on the mixed garnet separate from the whole rock is a mixed age, resulting from mixing of garnets from more

Table 4. Correlation of $^{207}\text{Pb}/^{206}\text{Pb}$ ages for Garnet and U–Pb ages for Zircon (in Ma) with rock types and metamorphic conditions

| Garnet ages | Zircon ages | Rock type and garnet and zircon separates | Metamorphic conditions |
|------------------|-------------------|---|---|
| Natawahunan Lake | | | |
| 2744–2738 | – | pelitic gneisses with garnets formed by breakdown of staurolite + quartz (B) and staurolite (C) | Upper amphibolite facies to granulite facies $T \leq 800^\circ \text{C}$ |
| Cauchon Lake | | | |
| ≥ 2984 Ma | – | Large anhedral garnet (A–K) in same graphic granite as garnet K | Amphibolite facies? |
| 2741 | – | Xenocrystic garnet (A–L) in same graphic granite as garnet L | Amphibolite facies? |
| 2700–2687 | 2695 ^a | Garnets from leucosome (D-1) and metapelite (E, F) corroded garnets in melanosome (M-1) Zircon from leucosome | Upper amphibolite facies with partial melting $T \geq 650$ |
| 2660–2637 | 2637 ^a | Garnets from metamorphosed pegmatite (G), metapelite (H) metarhyolite (I), and leucosome (M-1) Zircon from leucosome | Granulite facies with extensive $T = 750$ to 780°C |
| | 2629 | Zircon from E–W pegmatite (N) | |
| 2605–2591 | 2598 | Igneous garnets in undeformed leuco-granites (L, K) Zircon from N–S pegmatite (O) | Amphibolite facies intrusion of hydrous granites and pegmatites |

^a From Heaman et al. (1986); Krogh et al. (1986)

than one generation. The age of 2659 Ma for the single rounded garnet at the border between the leucosome and melanosome may be the time at which that type of garnet developed in the rock or it may also represent a mixed age.

The complexities in the garnets from these samples underscore the importance for using as small sample sizes as possible to avoid mixing of garnets grown at different times. Whenever possible individual grains should be selected which, based on petrographic observation, have formed by a single reaction. In some cases (samples A–K and K, A–L and L, D-1 and D2) even compositionally and petrographically very similar garnets may have formed at very different times.

U–Pb ages for Zircon and allanite

The three periods of garnet growth at 2700 to 2687 Ma, 2660 to 2637 Ma and 2605 to 2591 Ma from Cauchon Lake are similar to U–Pb ages determined for zircon and allanite from the same area (Table 1, 4):

2695 ± 3 Ma (Heaman et al. 1986a; Krogh et al. 1986)
 2637 ± 2 Ma (Heaman et al. 1986a; Krogh et al. 1986)
 2598 ± 1 Ma (Table 1, sample O)
 2600 ± 2 Ma (Table 1, sample P)

The zircon with the age of 2695 Ma is from a leucocratic mobilizate within a mafic gneiss. This mafic gneiss is adjacent to a pelitic gneiss containing granitic mobilizates with garnets. Garnet sample D-1 which gives an age of 2700 Ma is taken from this granitic mobilizate. The zircons with the age of 2637 Ma are taken from an amphibole-plagioclase mobilizate vein. The amphibole is very likely retrograde after orthopyroxene (T. Krogh and N. Machado, personal communication). This vein is exposed ca. 300 m west of the locality of migmatitic sample M in which garnet from the leucosome gives an age of 2641 Ma (Fig. 1).

As part of this study, U–Pb ages for zircon were obtained on small, metamict fragments of large single crystals (sample O ca. 1 cm, sample N ca. 0.3 cm) found in two structurally and mineralogically different late pegmatites. The east-west striking pegmatite (sample N) contains K-feldspar, quartz and biotite. The other pegmatite, striking north-south, (sample O) contains besides large zircons also K-feldspar, plagioclase, quartz, biotite, magnetite and allanite (Table 1).

For the U–Pb analyses pieces were taken from the cores of the stubby zircon crystals. Zircon O from the north-south striking pegmatite yields a concordant age of 2598 Ma, which is most probably the time of crystallization of this undeformed pegmatite. A completely metamict mineral (confirmed by XRD), probably allanite, coexisting with zircon O in the late pegmatite has a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2600 Ma and is ca. 3% reversely discordant (Table 1). Despite the discordance both the allanite and the zircon give $^{207}\text{Pb}/^{206}\text{Pb}$ and U–Pb ages which are in agreement. Recent disturbance of the U/Pb ratio may have caused the discordance of the allanite.

Zircon N from the east-west striking pegmatite shows reverse discordance of about 1%. Replicate analyses of the sample given the same discordance and the same $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2629 Ma ± 1 Ma. Reversely discordant zircons analyzed by conventional mass-spectrometric methods are rare (Davis and Trowell 1982). The reverse discordance is real and is not an artifact of the analytical technique. The blank correction is insignificant and the correction for inherited Pb is negligible because this zircon has a $^{206}\text{Pb}/^{204}\text{Pb}$ ratio of 30000–50000.

The K-feldspar in the pegmatite from which sample N was taken showed evidence of strong weathering. Thus, the zircon from this sample may have suffered recent U loss. Another explanation for the reverse discordance could be that some radiogenic Pb may have diffused from the rim toward the core of the zircon as found for reversely discordant zircons using the SHRIMP ion microprobe (Williams et al. 1984). The good agreement of the $^{207}\text{Pb}/^{206}\text{Pb}$ ages suggests that the pegmatite crystallized at 2629 Ma.

The overlap of the garnet and zircon ages indicates, that at Cauchon Lake garnet and zircon forming reactions occurred only in very specific time intervals, and both minerals can record the same large scale geologic events. In detail, however, garnet and zircon, which may form by different reactions, can give age information for different metamorphic conditions within a terrane.

Table 5. $^{207}\text{Pb}/^{206}\text{Pb}$ ages for garnet and pressure and temperature estimates made on the same sample

| Sample | $^{207}\text{Pb}/^{206}\text{Pb}$ age in Ma | P ^a kb | P ^b kb | T ^c °C |
|------------------|--|----------------------|----------------------|----------------------|
| Natawahunan Lake | | | | |
| B | 2744 | 7.1 | 9.5 | 800 |
| C | 2738, 2734 | 7.7 | 9.0 | 800 |
| Cauchon Lake | | | | |
| D | 2700, 2648 | – | 10.1 | 760 |
| E | 2694 | 7.3 | 6.7 | 780 |
| F | 2687 | 7.4 | 7.1 | 750 |
| H | 2646, 2644 | 7.0 | 8.4 | 750 |

^a Pressure estimated with equilibrium (1)

^b Pressure estimated with equilibrium (2)

^c Temperature estimated with two-feldspar geothermometer

Correlation of pressure, temperature and time

Table 5 gives the garnet $^{207}\text{Pb}/^{206}\text{Pb}$ ages as well as the metamorphic pressures determined on the same sample using equilibria (1) and (2). Also given are the temperatures which were determined on the sample used for dating, or for a sample from the same or an adjacent outcrop. Since the two-feldspar geothermometer is reset very easily, the temperatures most likely represent values for the last high grade thermal event at ca. 2640 Ma, or locally at ca. 2600 Ma. Because the U–Pb ages for garnet record the younger 2600 Ma or 2640 Ma thermal episodes and the older 2695 Ma, 2740 Ma or 2984 Ma episodes, it appears that the closure temperature for the U/Pb system in garnet is significantly higher than for the major elements Fe and Mg. As a consequence the pressures estimated with the garnet bearing equilibria are probably only reflecting the “peak” conditions during the last thermal event above 650°–700° C and not in all cases during the time of garnet growth. Therefore, in high-grade terranes information on the *P–T* conditions during the growth of the garnets has to be deduced from reaction textures. Only in terranes where the metamorphic conditions did not cause later re-equilibration of the major elements in garnet can the ages be directly correlated with the pressure and temperature of garnet growth as estimated by the geothermometers and geobarometers.

In Fig. 5, the U–Pb ages for the garnets which formed by a single reaction or partial melting are combined with temperature estimates to generate a *T–t* path. The prograde path (D-1, F and M-1 in Fig. 5) at Cauchon Lake is defined by reactions at 2700–2687 Ma and then later at 2660–2637 Ma. The metamorphism at 2700–2687 Ma attained at least upper amphibolite facies conditions as indicated by the formation of partial melts and the instability of the assemblage staurolite + quartz (Table 4). The metamorphism at 2660–2637 Ma reached granulite facies conditions and caused extensive partial melting. All the high two-feldspar temperatures and also the pressures determined from the various mineral equilibria are interpreted to represent the peak conditions reached during the metamorphic episode at ca. 2640 Ma.

The retrograde part of the path corresponds to the cooling following the metamorphism at 2660–2637 Ma. The in-

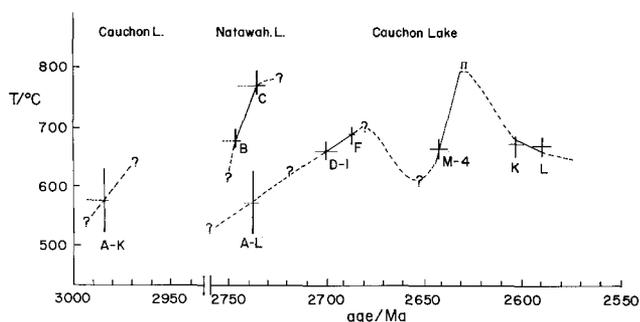


Fig. 5. This diagram shows possible temperature-time paths for the Pikwitonei-Cross Lake area based on $^{207}\text{Pb}/^{206}\text{Pb}$ garnet-K-feldspar ages. Based on xenocrysts of garnets in late granites, metamorphic conditions at Cauchon Lake were high enough at ca. 2984 Ma and ca. 2741 Ma (sample A-K and A-L) to cause the growth of garnet (minimum stability of garnet after Hsu 1968). At Natawahunan Lake the temperatures were above the stability field of staurolite + quartz (sample B) and staurolite (sample C) (Richardson 1968; Pigage and Greenwood 1982) at 2744–2738 Ma. At 2700 Ma partial melts of granitic composition were generated locally at Cauchon Lake (sample D-I); at 2687 Ma staurolite + quartz reacted to garnet + sillimanite + vapor (sample F). Extensive partial melting occurred at 2648–2637 Ma (samples D-2, M-4) and the last metamorphic garnets formed. The two-feldspar temperatures (labeled π) and probably the pressures shown in Fig. 1 and Table 5 represent the “peak” metamorphic conditions of this granulite grade metamorphism. Subsequently the terrane cooled and reached temperatures close to the minimum melting conditions for granites at ca. 2600 Ma (samples K and L). At this time granites and pegmatites intruded which caused the local retrogression of the granulite facies rocks in the southern part of the Cauchon Lake area to amphibolite grade and resetting of the two-feldspar temperatures (see Fig. 1b)

trusion of granites and pegmatites around 2600 Ma (K and L in Fig. 5) may be part of the same retrograde path. The intrusion of pegmatites and graphic granites at ca. 2629–2591 Ma, occurred at a time when the terrane may have cooled to temperatures near the minimum melting point of granite. The introduction of fluids together with these granitic melts in the southern part of the Cauchon Lake area near the amphibolite-granulite transition caused local retrogression of the granulite rocks to amphibolite grade and the resetting of the feldspar temperatures (Fig. 1b).

From petrographic data, the prograde metamorphism at Natawahunan Lake would be consistent with, and could reasonably be considered part of, the same prograde heating at Cauchon Lake. However, it can be seen (Fig. 5) that an upper amphibolite to granulite facies metamorphism at Natawahunan Lake occurred already at ca. 2744–2738 Ma. During this time metamorphic temperatures were sufficiently high to cause reaction of staurolite to garnet, sillimanite and hercynite. If a similar grade of metamorphism occurred at the same time in the Cauchon Lake area, it did not cause the formation of abundant garnet. A ca. 2741 Ma age is recorded at Cauchon Lake only in a xenocrystic garnet from a late granite.

Based on the large spread of the garnet ages one must conclude that the high grade metamorphism in this part of the Pikwitonei granulite domain lasted for at least 150 Ma. The agreement of the garnet and zircon ages may indicate that mineral reactions occurred only at very specific times within this period. The times of mineral growth may

be characterized by relatively rapid changes in pressure, temperature or fluid fugacities. The information recorded in the mineral assemblages of these rocks may thus in many cases reflect conditions which were attained only for a relatively short period of time during a long period of high grade metamorphism. Such “peak” metamorphic conditions were reached at different times in different areas (Table 4, Fig. 5).

Conclusions

Garnet is a major rock-forming mineral with U/Pb ratios that make it useful for dating. Garnets are closed to U–Pb diffusion up to at least 800°C, and therefore can record growth ages in most crustal rocks. Because of these features, age information deduced from garnets makes it possible to decipher the number of metamorphic episodes within a terrane as well as their duration and to date individual reactions. Previously, only zircon and rarely sphene (Tucker et al. 1987) were used to obtain high precision ages in complex metamorphic terranes. But the conditions for growth of zircon are not known, so that the zircon ages are not easily interpretable. Sphene, apatite and monazite can yield high precision ages, but are more easily reset (e.g. sphene: Hanson et al. 1971; Schärer 1980) and only rarely record multiple thermal events.

Pressure, temperature and time information coupled with thermal modelling will allow a quantitative reconstruction of the thermobarometric history of ancient mobile belts and may permit recognition of the style of tectonism, even for high grade terranes. So far, garnet has been widely used for obtaining metamorphic pressures and temperature. As shown by our study, it can also be an important mineral for obtaining high precision ages for metamorphic and igneous events. Such ages information will be critical to the development of quantitative thermal models for metamorphism.

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Appendix. Sample descriptions and locations (UTM coordinates):

Natawahunan Lake

B: metapelite with garnet-K-feldspar-plagioclase-biotite-quartz-sillimanite-cordierite-hercynite-rutile-ilmenite. The garnets are ≤ 0.5 cm in diameter (61771\6231).

C: paragneiss with leucocratic layers of K-feldspar-quartz-plagioclase and strongly peraluminous layers with garnet-sillimanite-hercynite-corundum-rutile-ilmenite and traces of biotite (61804\6175).

Cauchon Lake

A–K, K: coarse grained graphic granite with K-feldspar-plagioclase-quartz-biotite and schlieren enriched in garnet (ca. 0.1–1.0 cm) and biotite (61399\6527).

A–L, L: coarse grained leucogranite with schlieren enriched in garnet (ca. 0.2 to 0.8 cm) and biotite (61399\6527).

D: migmatite with a nebulitic texture. The melanosome consists of garnet-biotite-cordierite-spinel-K-feldspar-plagioclase-quartz-ilmenite, the leucosome contains garnet-K-feldspar-quartz-plagioclase. The size of the garnets is around 0.5 cm, single euhedral garnets for the leucosome were used for dating. (61547\6614).

E: metapelite with garnet-plagioclase-quartz-biotite-K-feldspar-cordierite-sillimanite-spinel-rutile-ilmenite-graphite. The garnets are ≤ 0.5 cm (61450\6443).

F: metapelite with garnet-K-feldspar-plagioclase-quartz-biotite-spinel-cordierite-ilmenite-rutile-graphite-monzonite-zircon. The garnets are ≤ 0.5 cm (61520\6562).

G: coarse grained, metamorphosed pegmatite consisting of plagioclase and quartz with traces of microcline and ca. 1% garnets (ca. 0.4 cm in diameter) (61523\6566).

H: metapelite with garnet-plagioclase-K-feldspar-quartz-sillimanite-cordierite-biotite-ilmenite-rutile. The garnets range from 0.2 to 0.8 cm (61459\6417).

I: metaryholite consisting of quartz-microcline with ca. 0.5% garnets (ca. 0.2 cm in diameter). Some of the garnets are rimmed by secondary biotite and show alteration to chlorite along fractures (61644\6724).

M: stromatic migmatite with garnet-biotite-quartz-plagioclase-K-feldspar-sillimanite in the melanosome and K-feldspar and garnet in the leucosome. In the melanosome the garnets range from 0.1 to 0.8 cm in the leucosome they are smaller than 0.2 cm (61395\6518).

N: K-feldspar-plagioclase-quartz-biotite-zircon pegmatite (61419\6500).

O, P: K-feldspar-plagioclase-quartz-biotite-magnetite-allanite-zircon pegmatite (61478\6483).