High-grade metamorphism in the Chapleau-Foleyet Area, Ontario

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Abstract

High-grade Archean rocks are exposed in the central Superior Province of the Canadian Shield in the Kapuskasing Structural Zone, a partly fault-bounded region up to 50 x 500 km. In migmatitic mafic gneiss, paragneiss, dioritic and tonalitic rocks, a lower-grade garnet–clinopyroxene–plagioclase (Gt–Cpx–Pl) zone and patchy, higher-grade orthopyroxene zones are distinguished. Grade decreases abruptly to greenschist in the Abitibi subprovince to the east across the Ivanhoe Lake cataclastic zone. Grade decreases gradually through amphibolite to greenschist facies in the Wawa subprovince to the west. Based on mineral-melt equilibria, minimum conditions for the Gt–Cpx–Pl zone are 750°C, 6 kbar, asr < 0.5–0.7 and for the orthopyroxene zone, 770°C, 6 kbar, asr < 0.5.

Various garnet–biotite and garnet–pyroxene geothermometers and geobarometers yield apparent temperatures ranging from <600°C in the west to locally >800°C. Apparent pressure values derived from the pyrope–grossular–anorthite–diopside–quartz equilibrium are 5.4–8.4 kbar (average 6.3) and define a NNE-trending area of relatively high P in the eastern and central Kapuskasing Zone, supporting the interpretation of a tilted crustal section.

Introduction

Regional metamorphism variably affected rocks of the Superior Province over an area of ~3 x 10^6 km². The erosion level throughout the southern part of the province exposes metamorphic rocks of the green schist and amphibolite facies except in the Kapuskasing Structural Zone, where pyroxene-bearing gneisses are associated with a linear gravity high (Innes, 1960; Bennett et al., 1967; MacLaren et al., 1968; Thurston et al., 1977). The Kapuskasing Zone in the relatively well-exposed Chapleau-Foleyet area consists of a variety of migmatitic pyroxene–garnet–hornblende–biotite-bearing rocks. The assemblages preserved at the time of metamorphic quenching provide insight into reactions leading to the decomposition of hornblende and biotite in an amphibolite–granulite facies transition zone. The metamorphic history and equilibration conditions of rocks in the Kapuskasing Zone and adjacent Wawa subprovince are examined in this study.

Metamorphism in the Kapuskasing Zone is Archean as inferred from U-Pb dates on zircon from leucosome in paragneiss (2627±5 Ma) and on metamorphic zircon in mafic gneiss (2650±2 Ma) (Percival and Krogh, 1983).

Geological setting

The Superior Province can be divided into metavolcanic-plutonic and metasedimentary subprovinces (Stockwell, 1970) (Fig. 1). In the south-central Superior Province, the easterly trends of the Abitibi-Wawa and Quetic-Opatica subprovinces are transected over a distance of 500 km by the north-northeast-trending Kapuskasing Structural Zone (Thurston et al., 1977) (Fig. 1). It is made up of gneisses in the upper amphibolite to granulite facies (Bennett et al., 1967; MacLaren et al., 1968) and is characterized by positive gravity and aero magnetic anomalies over most of its length (Innes, 1960).

The Chapleau-Foleyet area straddles the Kapuskasing Zone and includes parts of the adjacent Abitibi and Wawa subprovinces. Supracrustal rocks of the Abitibi subprovince in the study area comprise mostly mafic metavolcanic flows and tuffs up to 6100 m thick (Goodwin, 1965). Easterly-trending belts of felsic metavolcanic and metasedimentary rock make up about 5 and 3% respectively of the supracrustal succession. Massive to foliated plutons of tonalite to granite composition (Streckeisen, 1976) range in diameter from ~1 to ~25 km. The intensity of metamorphism increases from the cores of supracrustal belts, where green schist-facies assemblages are common (Thurston et al., 1977), to contact areas with large intrusive bodies, where hornblende–plagioclase±garnet assemblages prevail. Both primary structures, including bedding and volcanic features, and sub-vertical tectonic foliation, schistosity and lineation, are present in the supracrustal succession.

The contact between rocks of the Abitibi subprovince and the Kapuskasing Structural Zone to the west is the Ivanhoe Lake cataclastic zone (Fig. 2), comprising blan-
tomylonite, cataclasite and pseudotachylite veinlets (Percival and Coe, 1981). These fault rocks are sporadically developed in rocks of the eastern 2 km of the Kapuskasing Zone and have random orientation within individual outcrops. The overall NNE trend is deduced from the distribution of fault rocks in the field.

The Kapuskasing Zone consists of ENE-striking belts of paragneiss, mafic gneiss, tonalitic and dioritic rocks and units of the Shawmere anorthosite complex (Thurston et al., 1977; Percival, 1981a,b; Riccio, 1981). Paragneiss is layered, migmatitic, fine- to medium-grained, biotite-plagioclase-quartz rock, with some garnet and/or hornblende and/or orthopyroxene (Table 1). Concentrations of quartz, biotite, garnet, or graphite in some layers and the overall quartz-rich composition suggest that these rocks had a sedimentary origin. Mafic gneiss is a layered to homogeneous medium- to coarse-grained rock of high calcium (10.0-15.4 wt.% CaO), high-alumina (13.4-17.2 wt.% Al2O3) basaltic composition. It consists of Grt-Hb-Pl-Qz-Ilt assemblages and contains tonalitic leucosome veinlets on the 1-20 mm scale. Flaser diorite to mafic tonalite occurs as homogeneous to layered medium- to coarse-grained Hb-Pl±Qz±Cpx±Opx assemblages with up to 10% quartz and locally up to 15% concordant quartz monzonite veinlets. Rare gabbro and hornblendite layers may represent igneous differentiates. Tonalitic rocks are foliated to gneissic, locally xenolithic, consist of Bt-Pl-Qz±Hb±Gt assemblages, and contain paragneiss, mafic gneiss, and ultramafic (hornblendite or Gt-Opx-Hb rock) inclusions. Late hornblende-biotite tonalite dikes up to 50 cm thick transect gneissic layering in mafic and tonalitic gneiss. In 20-cm wide zones adjacent to these dikes, garnet, pyroxene-bearing assemblages are replaced by hornblende ± biotite assemblages. The Shawmere anorthosite complex (Fig. 2) comprises a main northern body, ~50 x 15 km, and a southern lens, ~15 x 4 km. The southern intrusion is homogeneous gabброng anorthosite whereas the larger body contains anorthosite through gabbro units as well as some ultramafic rocks. The northern body is crudely zoned, consisting of a thin marginal unit of migmatitic amphibolite, a layered anorthosite-gabbro unit and a core of plagioclase-megacrystic gabброng anorthosite (Riccio, 1981). A thin unit of foliated tonalite within the complex yielded a minimum U/Pb zircon date of 2765 Ma (Percival and Krogh, 1983) which provides a minimum age for the paragneiss-mafic gneiss country rock.

The western boundary of the Kapuskasing Zone is defined by the change in structural style and orientation from domal in the Wawa subprovince to northeasterly belts to the east (Fig. 2). Rare metaconglomerate in this boundary area has tonalitic cobbles with a U/Pb zircon date of 2664 ± 6 Ma and a K-Ar hornblende date of 2594 ±151 Ma (Percival et al., 1981).

The Wawa subprovince in the western part of the area (Fig. 2) is a tonalite to granodiorite gneiss terrane intruded by massive to foliated granite, quartz monzonite and tonalite plutons. The gneisses contain 0-25% amphibolitic enclaves as well as rare easterly-trending paragneiss units. Gneissosity is steep to subhorizontal, forming several domal structures, some with plutonic cores (Percival, 1981a). The Robson Lake dome, adjacent to the Kapuskasing Structural Zone (Fig. 2), has a core of mafic gneiss. Tonalite and granodiorite gneiss yield U/Pb zircon dates of >2707 and 2677 ±15 Ma respectively (Percival and Krogh, 1983).

The age of metamorphism in the Abitibi and Wawa subprovinces is constrained by zircon dates on deformed, metamorphosed volcanics of 2749-2696 Ma (Turek et al., 1982; Nunes and Pyke, 1980) and on post-metamorphic plutons of 2685-2668 Ma (Krogh and Turek, 1982; Krogh et al., 1982). Concordant U-Pb zircon dates of 2650 and 2627 Ma from high-grade rocks of the Kapuskasing Zone indicate either a discrete younger metamorphic event or sustained metamorphic effects of the pre-2685 Ma event (Percival and Krogh, 1983).

Metamorphic zones

Mineral assemblages in each lithologic unit are listed in Table 1 and plotted in Figure 3. Assemblage data in the Abitibi subprovince in this area are not sufficient for delineation of isograds. Chlorite-white mica-alkaline-quartz assemblages characterize metasediments and epidote-chlorite-albite-quartz-carbonate±white mica±hornblende assemblages are common in mafic rocks (Thurston et al., 1977). Hornblende-plagioclase±garnet assemblages occur in proximity to granitic plutons and locally near the Kapuskasing Zone.

Two high-grade metamorphic zones are recognized in the Kapuskasing Structural Zone and Wawa subprovince.
Fig. 2. Generalized geological map of the Chapleau–Foleyet area (modified after Percival (1981b)). Geological boundary between the Kapuskasing Zone and Wawa subprovince is gradational in this area. Approximate boundary is Chapleau River south to Robson Lake dome, then Nemegosenda River south to Lackner Lake complex.

The Gt-Cpx-Pl zone, with diagnostic assemblages in mafic gneiss, extends from the Ivanhoe Lake cataclastic zone in the east, westward into the Wawa subprovince. In addition to the diagnostic minerals, mafic rocks contain brown hornblende, quartz, ilmenite and/or sphene, rare scapolite and ubiquitous tonalitic leucosome veinlets comprising variable proportions of quartz and plagioclase with minor hornblende, garnet or clinopyroxene. Para-
Table 1. Mineral assemblages and densities of rocks in Kapuskasing Structural Zone and eastern Wawa subprovince

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Mineral abbreviations: Ol: olivine; Sp: sphene; Gt: garnet; Sl: sillimanite; St: staurolite; Sa: sapphire; Opx: orthopyroxene; Cpx: clinopyroxene; Hb: hornblende; Oa: orthoamphibole; M: muscovite; Bt: biotite; Pl: plagioclase; K: K-feldspar; Q: quartz; Sc: scapolite; I: ilmenite; Sn: spinel; Cc: calcite; Cu: cummingite; S.G.: specific gravity (bracketed numbers refer to average values); ± designation indicates that small quantities of the mineral are present as an additional phase in some assemblages.

gneiss from this zone generally has Gt–Bt–Pl–Qz assemblages with tonalitic leucosome consisting of quartz and plagioclase (An_{77-33}) with minor biotite. Only one sillimanite occurrence is known. Coarse muscovite is present in two localities near the margins of the Kapuskasing Zone, and one occurrence of staurolite–muscovite was noted (Fig. 3).

The orthopyroxene zone is represented by four areas with orthopyroxene-bearing rocks, surrounded by the Gt–Cpx–Pl zone. Isolated orthopyroxene localities and occurrences in anorthositic rocks, which may be igneous in origin, do not constitute orthopyroxene zones on Figure 3. In the orthopyroxene zone, small quantities of hypersthene occur, dominantly in paragneiss, but also in mafic and dioritic rocks. This implies that orthopyroxene forms in all rock types at similar grade, in contrast to the Adirondacks, where orthopyroxene forms in mafic rocks at lower grade than in felsic compositions (Buddington, 1963). In paragneiss, orthopyroxene, garnet and minor bleb antiperthite are locally present in leucosomes and orthopyroxene–K-feldspar–biotite occurs in melanosomes. Garnet, clinopyroxene and hornblende in specimen MG-20 are overgrown by fine symplectite of orthopyroxene–plagioclase, similar to the texture depicted by Horrocks (1980). This specimen contains three discrete plagioclase compositions: intergrowths of An_{32} and An_{39} occur in the matrix and An_{88} is present in symplectite.

Characteristic assemblages in lower-grade mafic rocks of the Wawa subprovince and Kapuskasing Zone are Hb–Pl–Qz±Sp±II and Hb–Cpx–Pl–Qz; the approximate distribution of each assemblage is indicated in Figure 3. Felsic gneiss is made up of assemblages of Bt–Pl–Qz±Hb±Ksp. One occurrence each of Gt–Opx–Bt and Opx–Cpx–Pl are known from paragneiss. Assemblages of Gt–Cpx–Hb–Pl–Qz are common in the Robson Lake dome.

The transition from the Cpx–Hb–Pl zone, characteristic of the eastern Wawa subprovince, to the Gt–Cpx–Pl zone is not sharp. Where mafic gneiss with Gt–Cpx–Hb–Pl assemblages occurs as enclaves in gneiss, the xenolith margins have hornblende-plagioclase assemblages. Therefore, what may have been a prograde sequence has apparently been subsequently retrograded by felsic intrusions or retrogressing solutions.

With the exception of some orthopyroxene-bearing rocks with chlorite veinlets and cataclastic rocks altered to chlorite–epidote, the minerals in most specimens are fresh and are therefore suitable for chemical study.

**Mineral chemistry**

All mineral analyses were carried out on an ARL model AMX microprobe equipped with an energy dispersive detector set up and supervised by P. L. Roeder and M. I. Corlett at Queen’s University. Operating conditions were
Fig. 3. Metamorphic mineral assemblages and index mineral isograds for part of the Chapleau-Foleyet area. Mineral abbreviations: Gt: garnet; Opx: orthopyroxene; Cpx: clinopyroxene; Hb: hornblende; Bt: biotite; Pl: plagioclase; Ksp: K-feldspar; Qz: quartz; ton: tonalitic segregations.

15 kV, 0.5 μA. Counting times of 120 seconds were used and data processed by an on-line minicomputer set up to apply Bence-Albee (1968) corrections and calculate weight per cent of oxides.

Between three and six spot analyses were made of each mineral in a probe section. Early detailed work showed that the average of three analyses is not significantly different from the average of six and therefore most of the results are averages of three analyses per mineral. Most minerals are homogeneous, with some chemical zonation within ~5 μm of the rim. Where rim composition is distinctly different from the interior analyses, it was excluded from the average.

Ten element analyses were recast into structural formulae according to the method of Deer et al. (1966). An estimate of Fe$^{3+}$ content of garnet and pyroxene was made by assuming electroneutrality and adjusting Fe$^{2+}$/Fe$^{3+}$ accordingly. Maximum and minimum Fe$^{3+}$ contents of hornblende were calculated by a method analogous to that of Stout (1972) and then averaged.
Mineral compositions from 25 paragneiss specimens are given in Appendix 1 and some compositional parameters are listed in Table 2. Garnet in paragneiss varies in Mg/(Mg+Fe) from 0.15 to 0.35. Grossular component constitutes 3-18 mol% and spessartine 1.1-11.7 mol%. Combined grossular-spessartine is generally 8-15 mol% with a few values as high as 25% (Table 2). Garnets are commonly zoned near the rim and the general pattern is toward increasing almandine (0-8 mol%) and spessartine (0-4 mol%) and decreasing pyrope (0-7 mol%) and grossular (0-3 mol%) components. The cores of grains are commonly homogeneous and the near-rim zonation may be a surface retrograde effect (cf. Lasaga et al., 1977). Orthopyroxene in paragneiss has a range of Mg/(Mg+Fe) of 0.44-0.62 and Al2O3 constitutes 0.9-3.5 wt%. Biotite is characterized by Mg/(Mg+Fe) = 0.35-0.68 and contains up to 4 wt% TiO2. Plagioclase is in the range An24-35 with one exception of An66. Analyses of coexisting garnet, orthopyroxene, biotite, K-feldspar and plagioclase from specimen PG-21 are presented in Table 3.

Mineral analyses from 34 mafic gneiss specimens are given in Appendix 1, and compositional parameters are listed in Table 4. Two of the specimens in the MG-series, MG-1 and MG-2, are skarns, but are included because of their grossular-rich garnet-clinopyroxene assemblages. Garnet in mafic gneiss is in the Mg/(Mg+Fe) = 0.13-0.43 range and grossular component constitutes 18-34 mol%. Spessartine is less than 5 mol% and andradite component is less than 5 mol%, an exception being MG-2, with 20 mol% andradite. Garnets generally have homogeneous interiors but are weakly zoned immediately adjacent to the rim. Spessartine and grossular contents vary only slightly whereas almandine increases (0-5 mol%) and pyrope decreases (0-4 mol%) within 5-10 μm of the grain boundary. Clinopyroxene is salite, with Mg/(Mg+Fe) values of 0.45-0.80. Jadeite and Ca-tschermakite components are in the 0-22 and 0-7.8 mol% ranges respectively. Amphiboles are common hornblende except for some secondary actinolite. Amphibole which appears to be a part of the equilibrium assemblage is brown and richer in Al2O3 (10-12 wt%) and TiO2 (0.4-2.2 wt%) than green amphibole of apparent retrograde origin (8-10 wt% Al2O3; 0.03-0.56 wt% TiO2). An exception is secondary sodic amphibole from MG-30, with 415 wt% TiO2. Plagioclase is andesine to labradorite (An35-50) and shows weak normal zoning. Two coexisting plagioclases in MG-20 (An35, An50) appear to be related by the Böggild miscibility gap (Ribbe, 1975). Ilmenite contains less than 0.5 wt% MgO. Scapolite has been noted in a few specimens, in one to the exclusion of plagioclase. Orthopyroxene is present in only two mafic gneiss specimens (MG-20, 22), and contains up to 4.7 wt% Al2O3. Analyses of coexisting garnet, orthopyroxene, and hornblende in specimen MG-20 are presented in Table 5.

Mineral analyses of dioritic, ultramafic and anorthosite-suite rocks are included in Appendix 1 and parameters listed in Table 6. Analyses of coexisting orthopyroxene, clinopyroxene, hornblende, biotite and plagioclase from a dioritic rock (OG-6) are presented in Table 7.

Graphical representation of assemblages

Minerals in mafic gneiss can be approximately represented in the system SiO2-Al2O3-FeO-MgO-CaO-Na2O-H2O, by ignoring the minor amounts of Fe2O3 calculated for the mafic phases. At fixed pressure, temperature and Na2O-H2O in the presence of quartz and plagioclase of constant composition, phase relations can be shown in terms of three components:

\[ A = Al_2O_3 - (CaO + Na_2O) \]
\[ F = FeO \]
\[ M = MgO \]

Mineral compositions can thus be plotted on a diagram using A/F+M and M/F+M coordinates (Stout, 1972).

Projections of coexisting hornblende, clinopyroxene, orthopyroxene and garnet define a crossing tie-line (Fig. 4). Consequently, the four minerals are stable along a line in a P-T diagram if \( a_{H_2O} = f(P,T) \); at lower temperature, hornblende and garnet are stable and at higher temperature, orthopyroxene and clinopyroxene are stable (Fig. 4). If water activity varies independently of \( P,T \), a decrease in \( a_{H_2O} \) at constant temperature would favor the orthopyroxene-clinopyroxene assemblage.

The assemblage garnet-clinopyroxene-hornblende is stable over a P-T range, represented by a shifting triangle on Figure 4. The garnet-clinopyroxene zone is reached when the bulk composition of the rock falls into the garnet-clinopyroxene-hornblende field. The orthopyroxene zone is reached when the path of metamorphism intersects the reaction

\[ \text{garnet} + \text{hornblende} \rightleftharpoons \text{orthopyroxene} + \text{clinopyroxene} \]  
(1)

The path of metamorphism extends only slightly beyond this reaction because the incompatibility of garnet-hornblende is never fully established.

Pressures and temperatures of migmatite formation

Tonalitic veinlets, consisting of plagioclase (An27-33), quartz and minor biotite, orthopyroxene, garnet, clinopyroxene or hornblende, occur in mafic rocks and in paragneiss as concordant layers and discrete pods. Several modes of origin are possible: 1) crystallization from...
injected melt; 2) metamorphic differentiation; or 3) crystallization from in situ anatectic melt. The injected melt hypothesis is improbable because the pods and veinlets are generally isolated. In addition, the mafic mineral content of the veinlets is the same as that in the host. Both characteristics suggest local derivation. Metamorphic segregation was invoked by Amit and Eyal (1976) to explain quartz-plagioclase leucosome in the Wadi Maghrish migmatites. This process is probable at sub-solidus temperatures (~630°C; Amit and Eyal, 1976), but is a more tenuous hypothesis for the present suite of rocks where indicated temperatures are much higher (see Geothermometry). Above 690°C at 5 kbar, in the presence of an aqueous fluid, which would be required to
transport the quartz and plagioclase components into the metamorphic segregation, quartz-plagioclase-H2O should form a melt phase (Yoder, 1967). The modal composition of tonalite veinlets in the Kapuskasing migmatites (35-50%) is consistent with melt compositions in the 700-800°C range (Yoder, 1967). Crystallization from in situ trondhjemitic melt was similarly invoked by Ashworth (1976) to explain the absence of K-feldspar in leucosome of biotite-quartz-plagioclase migmatites.

Experimental anatexis of biotite-quartz-plagioclase rocks produces a granite minimum-melt composition (Brown and Fyfe, 1970; Winkler, 1979). The source of KAlSi3O8 component must be biotite (with quartz), implying biotite instability at temperatures less than 700°C and the presence of other reaction products, such as amphibole or orthopyroxene. This is in disagreement with the refractory nature of biotite in biotite-quartz-plagioclase assemblages (Clemens and Wall, 1981).

In quartz-bearing mafic rocks, anatectic melts are tonalitic in the range 690-900°C at P_{H2O} = 5 kbar (Holloway and Burnham, 1972; Helz, 1976). The composition and proportion of plagioclase in tonalite veinlets from Kapuskasing mafic gneiss is consistent with an origin by crystallization from melt and therefore, a melt phase is considered to have coexisted with the solid phases at peak metamorphic conditions. By using the equilibrium relationship α_{H2O} in melt = α_{H2O} in vapor and assuming ideal mixing in the vapor and partitioning of H2O into the melt (Kilinc, 1979), one can calculate the shift of dehydration reaction curve (2) caused by equilibrium with undersaturated melt of quartz diorite composition (cf. Eggler, 1972):

4CaAl2Si2O8(OH)2 + 3CaAl2SiO8
in hornblende in plagioclase

⇒ 3Mg3Al2Si3O12 + 11 CaMgSi2O6 + 7SiO2 + 4H2O
in garnet in clinopyroxene quartz (2)

At any P,T, one can obtain:

ΔG° = -RT \ln \frac{α_{Py} \cdot α_{Cpx} \cdot f_{H2O}}{α_{Hb} \cdot α_{Frem} \cdot α_{Pl}} - ΔV_f(P - 1) (3)

Values of ΔG° and ΔV_f were obtained from Helgeson et al. (1978) and Haselton and Westrum (1980). The activity of Mg-components in minerals from four rocks spanning the range of compositional variation (x_{Mg} = 0.09-0.33) have been estimated from microprobe analyses and ideal ionic solution models. For the purpose of this calculation, a_{An} = X_{An}. By calculating the position of the curve for various values of α_{H2O}, one can estimate the P-T stability field of garnet-clinopyroxene-hornblende assemblages in the presence of tonalite melt (Fig. 5). The calculations were performed by an APL computer program written by D. M. Carmichael of Queen's University. Although there are large uncertainties in the position of the dehydration curve, resulting from uncertainties in ΔG° values, the calculated curve has a similar shape and position to experimentally-determined amphibole-breakdown curves for similar compositions, reported by Allen et al. (1975). The vapor-absent melting reaction within the divariant field is:

hornblende + plagioclase ⇒ garnet

+ clinopyroxene + tonalite (4)

Two intersections between dehydration and water-undersaturated melting curves occur for water activity values in the range 0.4-1.0; the minimum value occurs at about 3 kbar and water activity increases to 1.0 at 0.5 and at 15 kbar. As temperature increases beyond the curve, the proportion of products increases and water activity decreases. The overall negative slope of the vapor-absent hornblende melting curve is a function of the shapes of granitic solidi (concave toward high P,T) and amphibole dehydration curves (concave toward low P). These curves intersect at two P,T points where α_{H2O} is 1 by definition. The vapor-absent melting curve joining these two points has a negative slope, which is in contrast to the experimental vapor-absent melting curves for hornblende in acid and intermediate compositions of Brown and Fyfe (1970), which show strong positive dP/dT slopes. The

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<th>Table 3. Microprobe analyses of minerals in PG-21</th>
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1: garnet; 2: orthopyroxene; 3: biotite; 4: K-feldspar; 5: plagioclase

* Total iron as FeO; Fe^2+ by stoichiometry
Specimen also contains quartz and secondary chlorite
Table 4. Mafic gneiss mineral composition and equilibria data

<table>
<thead>
<tr>
<th>No.</th>
<th>Assemblage</th>
<th>Metamorphic Zone</th>
<th>Garnet</th>
<th>Clinopyroxene</th>
<th>Ti+Ct</th>
<th>Pl*bar</th>
<th>$n_{H_2O}$</th>
<th>Comments</th>
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<td>.024</td>
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<tr>
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<td>.061</td>
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<td>.692</td>
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<td>.436</td>
<td>.208</td>
<td>.023</td>
<td>.802</td>
<td>.049</td>
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</table>

A: Mineral abbreviations as in Table 1; dao Ma: magnetite; E: epidote
B: Metamorphic zones: 3: Gt-Cpx-Pl, 4: Opx
C: Molar fraction (X) of component i = U_i + F + K_i
D: $T^K = Mg/Mg + Fe in clinopyroxene
E: $X_{CaTs} = Al IV + (Fe + Mg + Fe in clinopyroxene)
F: $X_{Pl} = Ca + Ca + Na + K in plagioclase
G: based on Fe-Mg partitioning between garnet and clinopyroxene (Ellis and Green, 1979)
H: based on equilibrium (10) (Perkins and Newton, 1981)
I: based on equilibrium (3)
J: based on Fe-Mg partitioning between garnet and biotite (Ferry and Spear, 1978)
K: actinolite-plagioclase calibration (Goldsmith and Newton, 1978)
L: ilmenite-magnetite calibration (Buddington and Umbdey, 1964)
M: ilmenite-clinopyroxene calibration (Bishop, 1980)
N: garnet-orthopyroxene calibration (Wood, 1974)
Q: based on equilibrium (12), (13) (Perkins and Newton, 1981)
R: two pyroxene calibration (Powell, 1978)
The origin of many Archean tonalites has been ascribed to partial melting of mafic rocks containing garnet-clinopyroxene-quartz (Arth and Hansen, 1972) or garnet-clinopyroxene-orthopyroxene-plagioclase (Gower et al., 1982). Alternatively, dilution of the ambient fluid by externally-derived CO₂-rich vapors (Weaver, 1980; Friend, 1981; Janardhan et al., 1982) could induce crystallization and prevent retrogression, in the manner outlined by Wendlandt (1981).

Superimposed on Figure 6 is the vapor-phase absent melting curve for mafic rocks. The two curves intersect near 6 kbar, 775°C for natural compositions. This implies that in regional metamorphic terranes characterized by pressure above this intersection, hornblende will start to react to form garnet-clinopyroxene-melt at temperatures where paragneiss may be migmatitic but contains no orthopyroxene. In lower-pressure environments, biotite will become unstable at lower grade than hornblende, or hornblende may be involved in reactions producing orthopyroxene-bearing paragneiss.

The initial melt composition in the leucosomes is tonalitic but becomes granodioritic after K-feldspar is produced by decomposition of biotite, in accord with the observation of up to 10% antiperthite in leucosome of orthopyroxene-bearing paragneiss.

The curve (Fig. 6) has a steep positive dP/dT slope and assuming P of 6 kbar, orthopyroxene is stable above 770°C, in agreement with the experiments of Wendlandt (1981) and Clemens and Wall (1981).

The preceding discussion has assumed a closed system and equilibrium between melt and solids, however it is difficult to explain the presence of fresh orthopyroxene in tonalite veinlets if the system remained closed until crystallization. Reactions between water dissolved in the melt and orthopyroxene should yield hydrous phases. Hence water is considered to have been removed or diluted during or prior to crystallization. Removal could have been accomplished by collection and upward migration of tonalitic liquids leaving a water-depleted residue. The origin of many Archean tonalites has been ascribed to partial melting of mafic rocks containing garnet-clinopyroxene-quartz (Arth and Hansen, 1972) or garnet-clinopyroxene-orthopyroxene-plagioclase (Gower et al., 1982). Alternatively, dilution of the ambient fluid by externally-derived CO₂-rich vapors (Weaver, 1980; Friend, 1981; Janardhan et al., 1982) could induce crystallization and prevent retrogression, in the manner outlined by Wendlandt (1981).

Superimposed on Figure 6 is the vapor-phase absent melting curve for mafic rocks. The two curves intersect near 6 kbar, 775°C for natural compositions. This implies that in regional metamorphic terranes characterized by pressure above this intersection, hornblende will start to react to form garnet-clinopyroxene-melt at temperatures where paragneiss may be migmatitic but contains no orthopyroxene. In lower-pressure environments, biotite will become unstable at lower grade than hornblende, or hornblende may be involved in reactions producing orthopyroxene in mafic rocks (cf. Wells, 1979).

Minimum pressure-temperature conditions were de-
Table 6. Orthogneiss mineral composition and equilibria data

<table>
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<tr>
<th>No.</th>
<th>Assemblage A</th>
<th>Metamorphic Zone</th>
<th>Gt</th>
<th>Cpx</th>
<th>Opx</th>
<th>Pl</th>
<th>Hb</th>
<th>Sn</th>
<th>Ti&lt;sup&gt;4&lt;/sup&gt;</th>
<th>P(bkar)&lt;sup&gt;3&lt;/sup&gt;</th>
<th>a&lt;sub&gt;H2O&lt;/sub&gt;</th>
<th>Comments</th>
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<tr>
<td>OG-1</td>
<td>Gt Sa Opx Hb Oa Pl Sn</td>
<td>3</td>
<td>0.561</td>
<td>0.09</td>
<td>0.81</td>
<td>0.098</td>
<td>0.9%</td>
<td>(750)</td>
<td>5.3</td>
<td>Anorthosite-suite Melagabbro</td>
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<td>OG-2</td>
<td>Opx Cpx Hb Bt Pl</td>
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<td>0.80</td>
<td>0.038</td>
<td>0.92</td>
<td>0.9%</td>
<td>(750)</td>
<td>7.0</td>
<td>0.01</td>
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<td>OG-3</td>
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<td>0.592</td>
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<td>0.82</td>
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<td>0.9%</td>
<td>(750)</td>
<td>7.0</td>
<td>Anorthosite-suite Melagabbro</td>
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<td>(750)</td>
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<td>OG-7</td>
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<td>0.870</td>
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<td>0.05</td>
<td>Mg-rich Ultramafic; T&lt;sup&gt;4&lt;/sup&gt; = 800°C</td>
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A: Mineral abbreviations as in Table 1  
B: Metamorphic zones: 1: Gt-Cpx-Hb; 2: Gt-Cpx-Hb-Bt; 3: Gt-Cpx-Hb-Bt Pl; 4: Opx Cpx-Hb-Bt Pl I Ca  
C: \( x_{Gt}^{Gt} = \frac{Mg}{Mg + Fe^2} \) garnet  
D: \( x_{Gt}^{Ctp} = \frac{(Cu/Ca + Mn + Mg + Fe^2)}{Fe^2} \) garnet  
E: \( x_{Al}^{Opx} = \frac{Mg}{Mg + Fe^2} \) orthopyroxene  
F: \( x_{Al}^{Opx} = \frac{(Al + Ca + Fe^2 + Mg)}{Fe^2} \) orthopyroxene  
G: \( x_{Cpx}^{Cpx} = \frac{Mg}{Mg + Fe^2} \) clinopyroxene  
H: \( x_{Al}^{Pl} = \frac{(Ca + Na + K)}{Pl} \) plagioclase  
I: Based on Powell's (1978) two-pyroxene thermometer. Bracketed values are assumed for the pressure calibration  
J: Based on Wood's (1974) aluminous orthopyroxene-garnet barometer. Bracketed values are assumed for the temperature calibration  
K: Based on equilibrium (13)  
L: Based on Roeder et al. (1978) olivine-spinel thermometer

fined in the previous section by solid–melt equilibria for the Gt–Cpx–Pl zone and for the orthopyroxene zone. Regional variation in apparent P–T conditions is now examined, based on calibrated geothermometers and geobarometers applied to individual samples. (Tables 2, 4 and 6)

**Temperature estimation methods**

Geothermometers based on Fe–Mg exchange between mineral pairs are sensitive to temperature variation and only slightly pressure-dependent. Therefore, to evaluate temperature-variation on a first approximation basis, a value of 7 kbar was initially estimated for all specimens. This value cannot be so precisely estimated but is shown below to be a reasonable average for the area.

The mineral pairs most suitable for geothermometry by virtue of widespread distribution and unaltered character are garnet–biotite and garnet–clinopyroxene. The Ferry and Spear (1978) experimental calibration of the garnet–biotite thermometer has gained general acceptance (e.g., Ghent et al., 1979; Ganguly, 1979), the experimental calibration of Ellis and Green (1979), which takes account of the effect of calcium on Fe–Mg distribution, is well-suited for the study of the Kapuskasing Zone mafic rocks whose garnets contain up to 34 mol% grossularite. Mineral compositions meet the compositional restrictions imposed in both garnet–biotite and garnet–clinopyroxene calibrations (Appendix 1).

Figure 7 shows temperature estimates of the Ferry and Spear calibration plotted against estimates by the three most recent garnet–clinopyroxene calibrations for four rocks from the Kapuskasing Zone that contain all three minerals. Internal consistency is best for these samples for the Ellis and Green and Ferry and Spear calibrations. Both are experimentally-derived equations and take account of pressure effects. The agreement between the two techniques is in contrast to that reported for Adirondack granulites studied by Bohlen and Essene (1980).

Stoichiometric estimation of Fe<sup>3+</sup> in biotite from probe data is not possible because site vacancies in biotite leave the structural formula cation-deficient. However, wet chemical analyses of biotite consistently indicate the presence of Fe<sub>2</sub>O<sub>3</sub>. To assess the partitioning of Fe<sup>3+</sup>
ed Fe$^{2+}$ value is used to calculate $K_D$. Further justification for using total Fe rather than Fe$^{2+}$ comes from samples MG-17 and MG-25, whose garnet and clinopyroxene contain no Fe$^{3+}$ as calculated stoichiometrically. Temperatures of 725 and 755°C are in the same range as values calculated by ignoring Fe$^{3+}$ in other specimens (Table 4).

A calibration of the two-pyroxene thermometer, proposed by Powell (1978), yields results which are consistent with garnet-biotite and garnet-clinopyroxene temperatures for the Kapuskasing gneisses. The equations formulated by Kretz (1982) give averaged values in the same range but some pyroxene pairs yield temperatures that are up to 100°C discordant. The Wood and Banno (1973) and Wells (1977) calibrations consistently give temperatures some 200–300°C higher.

Additional results were obtained for rocks containing mineral assemblages without widespread distribution (Tables 2, 4 and 6). In general, oxide, feldspar and ilmenite-pyroxene thermometers yield temperatures some 200–300°C lower than those estimated by the Fe-Mg exchange thermometers. No generalizations regarding other thermometers are possible because of the limited number of occurrences.

**Regional temperature variation**

Near-rim garnet analyses generally give results lower by 10–20°C than analyses of the relatively homogeneous interiors. The general analytical strategy was to check all

![Fig. 4. Phase relations of mineral assemblages coexisting with tonalitic melt in mafic rocks. Mineral compositions are generalized from Appendix I for a and c; plotted from Table 5 for b. (a) garnet-clinopyroxene–hornblende compositional triangles shift to the right along M/M+F axis in response to rising temperature; (b) garnet + hornblende react to produce orthopyroxene + clinopyroxene with increasing temperature (or decreasing $a_{H_2O}$ if $P$, $a_{H_2O}$ are independent of $T$); (c) stable mineral assemblages in the orthopyroxene zone.](image-url)
the area as well as isolated occurrences in the 800°C range near the southeastern margin of the Kapuskasing Zone. Most of the Gt-Cpx-Pl zone is characterized by values above 700°C. Rocks which give apparent lower temperatures, in the 600°C range, are in the southwest corner of the area, where garnet-, biotite-bearing paragneiss schlieren occur as inclusions in tonalitic gneiss.

A high paleotemperature zone is present near the northern end of the Shenango complex (Fig. 2) and a possibly similar relationship exists east of the Nemegosenda Lake complex. The origin of the apparent thermal highs may relate to the late (1100 Ma) intrusive bodies. Samples from the vicinity of the Nemegosenda complex may have been affected by contact metamorphism, metamatism, or crustal buckling or displacement adjacent to the body. Any of these processes could alter the apparent temperature of adjacent rocks, the first by readjusting Fe–Mg ratios of garnet and clinopyroxene, the second by increasing Ca content of garnet, and the third by displac-

grains for zoning but exclude near-rim analyses from the average.

The results of the combined garnet–biotite, garnet–clinopyroxene and two-pyroxene thermometers are plotted in Figure 8, which shows an overall temperature range of 580–825°C. The extremes in this range are garnet–biotite temperatures; garnet–clinopyroxene results have a narrower spread (655–815°C). Much of the area without data points is underlain by diorite or tonalite orthogneiss without relevant assemblages.

The apparent temperature distribution of Figure 8 indicates two zones above 800°C in the northern part of

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Fig. 5. Vapor-phase-absent-melting curve (diagonal ruling) corresponding to the reaction Hb + Pl = Gt + Cpx + tonalite for mafic rocks of variable Mg/(Mg+Fe) ratio. The curve is defined by the locus of intersection between dehydration (Reaction 2; dot-dash symbol) and melting (Kilinc, 1979) curves of equal water activity. Position of curves for reaction Opx + Pl = Gt + Cpx + Qz at variable XGt are based on Hansen (1981; p. 239) Symbols as for Fig. 3 except Py: pyrope; Tr: tremolite; Di: diopside; An: anorthite.

---

Fig. 6. Vapor-phase-absent-melting curve for paragneiss. Construction as for Fig. 5. Stippled area is VPAM curve for mafic rocks. Note that the two VPAM curves intersect at 6–8 kbar, 775°C for natural compositions.
Fig. 7. Comparison of temperature estimates derived from garnet-biotite and various garnet-clinopyroxene thermometers. The two thermometers are consistent if the Ferry and Spear (1978) and Ellis and Green (1979) calibrations are used and $K_p$ is calculated using the total iron content rather than stoichiometrically-determined $Fe^{2+}$.

Fig. 8. Temperature map of part of the Chapleau-Foleyet area. Temperatures estimated by garnet-biotite (Ferry and Spear, 1978), garnet-clinopyroxene (Ellis and Green, 1979) and two-pyroxene (Powell, 1978) thermometry. Numbers to the left of dash are sample identifiers; numbers to the right are in °C. Circles: PG-series; squares: MG-series; triangles: OG-series.
exsolution lamellae in metamorphosed Adirondack anor- 

thosite massifs as being of igneous derivation. They 

derived temperatures both of igneous crystallization 

(-1100±100°C) and metamorphism (-750°C) by reinte-

gration of pyroxene compositions. The same treatment 

cannot be applied to the Shawmere metagabbro because 

orthopyroxene is homogeneous and plots in the metamor-

phic temperature range on Ross and Huebner’s (1975) 

pyroxene solvus isotherm diagram. Assuming, however, 

that the exsolved clinopyroxene megacrysts at one time 

coexisted with a calcium-richer orthopyroxene, a tem-

perature on the order of 1050°C can be estimated for 

igneous crystallization. Based on the composition of 

matrix clinopyroxene and homogeneous orthopyroxene, 

an estimated temperature of metamorphism of 750°C may 

be obtained from the Ross and Huebner isotherm plot.

Pressure estimation methods

Pressure estimates are dependent on the temperatures 

derived in the previous section. Several pressure-sensi-

tive equilibria are relevant to assemblages with wide-

spread distribution in the Kapuskasing Zone. The com-

positions of minerals in garnet-pyroxene-plagioclase-

quartz parageneses are determined by equilibria, including:

\[
\begin{align*}
\text{CaAl}_2\text{Si}_2\text{O}_8 & \rightleftharpoons \text{CaAl}_2\text{Si}_6\text{O}_{12} + \text{SiO}_2 & (8) \\
2\text{CaAl}_2\text{Si}_3\text{O}_{12} + \text{Mg}_3\text{Si}_2\text{O}_{12} & \rightarrow 3\text{CaMgSi}_2\text{O}_6 + 3\text{CaAl}_2\text{Si}_6\text{O}_{12} & (9) \\
\text{CaAl}_2\text{Si}_2\text{O}_8 + \text{CaMgSi}_2\text{O}_6 & \rightleftharpoons 2/3\text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12} + \text{SiO}_2 & (10) \\
\text{CaAl}_2\text{Si}_2\text{O}_8 + \text{Mg}_2\text{Si}_2\text{O}_6 & \rightleftharpoons 1/3\text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12} + 2\text{Mg}_3\text{Si}_2\text{O}_{12} + \text{SiO}_2 & (11) \\
\text{Mg}_2\text{Si}_2\text{O}_6 + \text{MgAl}_2\text{Si}_6\text{O}_{12} & \rightarrow \text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12} & (12)
\end{align*}
\]

Anorthite activity in plagioclase was estimated by the 

Al-avoidance model of Perkins and Newton (1981). For 

MG-20, a visual estimate of the proportions of An$_{80}$ and 

An$_{12}$ plagioclase was made to estimate a weighted average of An$_{45}$. Diopside and enstatite activities are derived from 

data of Wood and Banno (1973). Interaction parameters 

were used to estimate pyrope and grossular activity in 

garnet according to Perkins and Newton (1981).

Geobarometers based on (10) and (11) have recently 

been calibrated by Perkins and Newton (1981); results 

from mafic gneiss of the Kapuskasing Zone are presented 

in Table 4. Wood (1977) applied Equilibrium (8) to 

granulites as a geobarometer. Using a calibration of (8) 

based on thermochemical data of Helgeson et al. (1978), 

pressure estimates are seen to be controlled mainly by the 

formulation of the CaAl$_2$Si$_6$O$_6$ activity model. Wood’s 

(1978) formulation leads to a cluster of values in the 4 

kbar range whereas Wood's (1979) model produces re- 

sults averaging ~15 kbar. Equilibria (9) and (12) are 

subject to the same uncertainty.

Wood’s (1974) calibration of (12) gives additional pres- 

sure estimates for garnet, orthopyroxene-bearing assem- 

blages. Pressure was estimated for a Gt-Sl-Pl-Bt-Qz 

assemblage according to Ghent’s (1976) calibration.

The accuracy of the various geobarometers is limited 

by the uncertainty in probe analyses, greatest where 

accurate analyses of small quantities of alumina in pyrox- 

ene are required ((8), (9), (12)). The low dPdT slope of 

(10) and (11) and their use of major components recom- 

mend these equilibria as reliable geobarometers.

Pressure estimates

Estimates of peak metamorphic pressure depend on the 

assumption that present mineral compositions are not 

significantly different than those at the peak metamorphic 

conditions. In samples with retrograde amphibole, this 

assumption may not be valid because other minerals, 

including plagioclase, could have changed their composi- 

tions during retrograde reactions, presumably at lower 

P,T. Such reactions could also have occurred in response 

to changes in a$_{17r6}$ at peak P,T.

The Perkins and Newton (1981) calibration of (10) 
yields pressure values generally in the range 5.4–8.4 kbar, 

averaging 6.3 kbar. The results are plotted on Figure 9 

along with values derived from (11) and (12). The average 

pressure for the area from (11) is 7.7 kbar and individual 

values are commonly ~2 kbar higher than those estimat- 
ed from (10) for the same or proximal samples. Values 

from (10) are generally lower than those from (11), either 
because of different blocking temperatures in different 

mineral systems or uncertainty in the thermodynamic 
calibration (Newton and Perkins, 1982). A 6.3 kbar refer- 
ence line separates areas with above-average apparent 

values from those with lower values. The line includes a 

roughly north-south-trending area of relatively high ap-

parent pressure in the central and eastern Kapuskasing 

Zone (Fig. 9). This pattern supports the preliminary 

interpretation (Percival, 1981a) that the Kapuskasing 

Zone is at the base of a tilted, west-dipping crustal 

section. However, diffusion considerations may preclude 

this simple interpretation (see below).

Assuming that pressure is a function of depth, approxi-

mately 20–25 km of overburden has been eroded from the 

Kapuskasing zone. Previous estimates of relative uplift of
Fig. 9. Paleopressure map of part of the Chapleau-Foleyet area. Symbols represent rock type (circles: PG-series; squares: MG-series; triangles: OG-series), keyed in to sample identifier (numbers to left of dash). Numbers to the right of the dash are pressure estimates, in kbar, keyed in to the equilibrium used to derive the value. The 6.3 kbar reference line is based on the anorthite-diopside-grossular-pyrope-quartz equilibrium.

the Kapuskasing Zone with respect to adjacent regions where 8 km, based on heat-flow data (Cermak and Jessop, 1971) and at least several thousand feet (Thurston et al., 1977). Pressure-sensitive assemblages are lacking in the terranes to the east and west. However, slates some 5 km east of the Ivanhoe Lake cataclastic zone is at relatively low grade (chlorite-muscovite). Although \( \frac{dP}{dT} \) is generally not constant with depth (P. H. Thompson, 1977), assuming an average crustal geothermal gradient of \(-30^\circ C/km\), biotite should form at the expense of muscovite-chlorite at about 450°C, corresponding to a depth of about 15 km. Therefore, relative uplift along the Ivanhoe Lake cataclastic zone is estimated to be at least 5 and possibly greater than 10 km, based on metamorphic considerations.

**Water activity calculations**

Water activity was calculated for rocks containing assemblages corresponding to Mg end-member dehydration equilibria for which \( P-T \) estimates were available. Three equilibria have been used for this purpose, including (2):

\[
Ca_2Mg_5Si_2O_{22}(OH)_2 \rightleftharpoons 3MgSiO_3 \\
\text{in hornblende in orthopyroxene} \\
+ 2CaMgSi_2O_6 + SiO_2 \text{ in clinopyroxene quartz} \tag{13}
\]

\[
KMg_3AlSi_3O_{10}(OH)_2 + 3SiO_2 \text{ in biotite quartz} \]

\[
\Delta G = -RT \ln K - \Delta V(P - 1) \tag{15}
\]

Activity of solids were estimated by ideal ionic solution models and thermochemical data are from Helgeson et al. (1978) and Haselton and Westrum (1980). Water activity is calculated as \( f_{H_2O} \) as defined by (15), divided by \( f_{H_2O} \) of pure water at \( P,T \) as defined by Helgeson and Kirkham (1974). Derived values are found in Tables 2, 4 and 6.

Equilibrium (13) yields values <0.1; (14) gives results <0.4 and (2) yields some values above 1.0. Values in excess of 1.0 indicate that the minerals were not in equilibrium at the specified \( P-T \) conditions. For example, brown hornblende in MG-11 yields \( a_{H_2O} = 0.853 \) whereas actinolite, presumed to be of retrograde origin, yields a mechanically-un sustainable value of 5.5. Green, secondary amphiboles indicate higher \( a_{H_2O} \) in every example (e.g., MG-4, 11, 20), but would yield lower \( a_{H_2O} \) values if a lower temperature were assumed.

No pattern of regional variation in water activity is obvious from values plotted on a map. Values are generally on the order of 0.1 near the western edge of the Kapuskasing Zone but do not show a regional trend for the rest of the area. There is no correlation between calculated water activity and the presence or absence of orthopyroxene. The estimates are in the range of values reported for granulite-facies terranes (e.g., Bohlen and Essene, 1976b) for the Adirondacks; Horrocks (1980) for the Limpopo Belt) and for the Broken Hill area of Australia (Phillips, 1980).

**Significance of pressure-temperature results**

A comparison between \( P-T \) estimates for individual samples and probable \( P-T \) conditions in the Gt-Cpx-Pl and orthopyroxene zones shows that temperature estimates for rocks containing orthopyroxene are below the Gt-Cpx zone conditions deduced by mineral-melt equilibria. Eight of twenty-five estimates from the Gt-Cpx-Pl zone fall within the Gt-Cpx zone on the \( P-T \) diagram; the rest fall below. This pattern suggests that Fe-Mg exchange between mineral pairs continued during cooling and that derived temperature estimates are somewhat retrograde.

For the Conemaqua region, Yardley (1977) suggested that temperature above 640°C had homogenized prograde zonation in garnets by intragranular diffusion. Lasaga et al. (1977) estimated that retrograde Fe-Mg exchange between the outer 10-15 µm of adjacent garnet and cordierite grains occurred down to \(-450^\circ C\). In view of the quenching problem it may be realistic to regard the highest apparent temperatures, in the \( 800^\circ C \) range, as closest to peak temperature. The lower values probably record quenching at later times. By this reasoning, the pressure and temperature estimates, based on mineral
systems with probable different blocking temperatures (O'Hara, 1977), may not even record a $P-T$ condition through which the rock passed. Thus it is hazardous to attempt to deduce either a prograde or cooling metamorphic path for the area, based on geothermobarometry.

**“High-pressure granulites”**

The assemblage almandine garnet–clinopyroxene–plagioclase–quartz is diagnostic of the regional hypersthene zone according to Winkler (1979, p. 260, 267–268). de Waard (1965) and Green and Ringwood (1967) suggested that this assemblage forms as an alternative to orthopyroxene–plagioclase during high-pressure granulite-facies metamorphism. Turner (1981) attaches a different significance to the assemblage, regarding it as transitional from amphibolite to granulite facies based on Binns’ (1964) study. In the present study area, the location of the Gt-Cpx-Pl zone between Hb-Pl±Cpx rocks and orthopyroxene-bearing rocks suggests that it characterizes the amphibolite–granulite facies transition. Although the assemblage is the same as that in the Adirondacks (de Waard, 1965) and temperature conditions were similar (Bohlen and Essene, 1977), the path of metamorphism was different. In the Grenville Province, the development of garnet-clinopyroxene assemblages has been attributed to isobaric cooling of orthopyroxene–plagioclase granulites (Martingole and Schrijver, 1971; Whitney, 1978) whereas in the Kapuskasing Zone, garnet and clinopyroxene formed during prograde reactions.

**Conclusions**

In the eastern Wawa subprovince and Kapuskasing Structural Zone, three metamorphic zones are defined on the basis of assemblages in mafic rocks. The sequence Hb-Pl, Cpx-Hb-Pl, Gt-Cpx-Pl suggests easterly-increasing grade. Both hornblende- and biotite-bearing rocks in four areas surrounded by the Gt-Cpx-Pl zone contain metamorphic orthopyroxene, representing the highest grade attained. Tonalitic melt is thought to have been responsible for reducing $q_{H_2O}$ in the metamorphic fluid in mafic gneiss to values of $0.5-0.7$ at temperatures of $750^\circ C$, at which point Hb-Pl began to produce Gt-Cpx melt assemblages by vapour-absent melting reactions. Biotite in adjacent migmatitic paragneiss units was stable at these conditions but began to break down to orthopyroxene-K-feldspar-melt by vapour-absent-melting reactions at $770^\circ C$, $q_{H_2O} \sim 0.5$. Orthopyroxene first appears in mafic rocks at similar conditions.

Estimates of metamorphic pressure and temperature by various calibrations are 3.1–9.9 kbar, 600–825°C but most values are between 6.0 and 7.3 kbar and 650 and 775°C. The orthopyroxene-zone rocks yield average to below-average values suggesting that the peak-metamorphic distribution of elements has been altered, probably by ionic diffusion during cooling.

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**References**


Arth, J. G. and Hanson, G. N. (1972) Quartz diorites derived by partial melting of eclogite or amphibolite at mantle depths. Contributions to Mineralogy and Petrology, 37, 161–174.


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