

Comparative geothermometry based on major-element and oxygen isotope distributions in Precambrian metamorphic rocks from southwestern Montana¹

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Abstract

Mineral assemblages in the Kelly and Carter Creek areas of the Ruby Range, southwestern Montana, indicate orthopyroxene-zone and sillimanite-K feldspar-zone metamorphism, respectively. Because the two areas are small and contain diverse lithologies, they are ideal for a comparative study of mineral-pair geothermometers and grade indicators. For the Kelly area, various major-element methods (clinopyroxene–orthopyroxene, garnet–orthopyroxene, garnet–hornblende, garnet–biotite, and K feldspar–plagioclase) yield a consensus peak temperature estimate of $745\pm 50^\circ\text{C}$. For the Carter Creek area, garnet–clinopyroxene, garnet–hornblende, garnet–biotite, and K feldspar–plagioclase methods plus assemblage data yield $675\pm 45^\circ\text{C}$.

Coexisting quartz and magnetite in metamorphic iron-formations from the two areas yield anomalously low temperature estimates of $530\pm 15^\circ$ and $475\pm 25^\circ\text{C}$, respectively, based on oxygen isotopic fractionations of 8.41 ± 0.30 per mil and 9.92 ± 0.20 per mil. Similarly, Mg distributions between calcite and dolomite in metadolomites from the Kelly area yield estimates of *ca.* 545°C . These lower temperatures are probably the result of (1) solid-state diffusion of chemical species between minerals or (2) recrystallization of quartz, magnetite, and carbonates (*i.e.*, the non-refractory minerals) during retrograde or subsequent low-grade metamorphism. Another low estimate (545°C) is obtained from rim compositions of coexisting garnet and cordierite in one magnesian meta-pelite sample from the Kelly area. This estimate is consistent with the observation of cordierite coronas on garnet, a texture that suggests retrograde origin for the cordierite.

These data show it cannot be assumed *a priori* that a given geothermometer will record peak metamorphic temperature. Instead, the bimodal discordance of estimates presented here reflects the varied response of different mineral pairs to a given metamorphic and tectonic history.

Introduction

Many published geothermometers are based upon the temperature-dependence of major-element or isotope distributions between coexisting minerals. An inherent danger confronting the petrologist who uses such methods is the tendency to regard calculated temperatures as estimates of *peak* thermal conditions in an area, without due consideration for possible alternatives. The primary purpose of this communication, therefore, is to evaluate the precision, accuracy, sensitivity, petrologic significance, and limitations of selected mineral-pair geothermometers.

The evaluation is based upon a comparison of temperature estimates inferred from geothermometers and mineral assemblages in two areas within the Ruby Range, southwestern Montana. These two areas, designated in Figure 1 as the Kelly and Carter Creek areas, are particularly well-suited to the proposed study because of their lithologic variety, small size, and differing metamorphic grades.

Geologic setting

The Ruby Range is a northeast–southwest-trending block whose highly deformed and metamorphosed Precambrian (Pre-Beltian) core was exposed by Laramide uplift and subsequent erosion. Sillimanite–K feldspar-zone metamorphism prevailed

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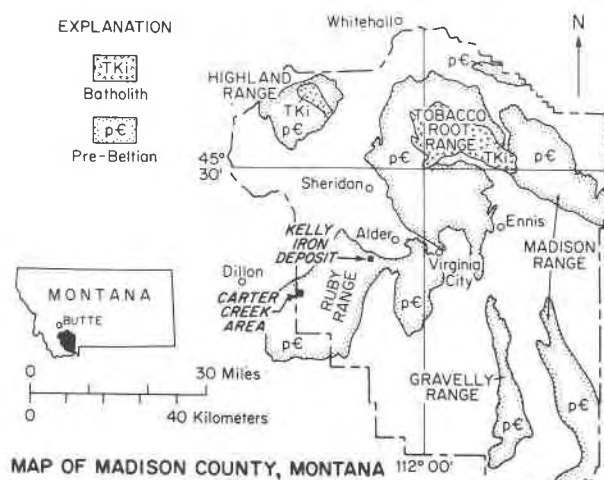


Fig. 1. Map showing exposures of Precambrian metamorphic rocks and Tertiary intrusives in Madison County, Montana (after Cordua, 1973). The Kelly and Carter Creek areas of the Ruby Range are the primary sample sites for this study. These areas are named for Precambrian metamorphic iron-formations contained therein. One sample (sample RMS-22) is from the Stone Creek area, located approximately half-way between the two main areas.

throughout most of the range, including the Carter Creek area (ca. 9 km² in area), as evidenced by mineral assemblages in pelitic rocks. However, in the northeast part of the area, which includes the Kelly iron deposit (ca. 1 km² in area), assemblages in mafic gneisses and iron-formations indicate that the orthopyroxene isograd was reached. Mineral assemblages in the two areas are summarized in Table 1.

Both areas are situated within a thick sequence of metamorphosed dolomites, mafic gneisses, iron-formations, pelitic and semi-pelitic rocks, and quartzites. These rocks, collectively termed "Cherry Creek" rocks by Heinrich (1960), are thought to have attained their high-grade metamorphic character approximately 2700 m.y. ago, during the Beartooth orogeny (H. L. James, personal communication). The "Cherry Creek" rocks are conformably underlain by the Dillon pelitic gneiss unit, which in turn is underlain by a sequence of "pre-Cherry Creek" schists and gneisses. The regional foliation of the Pre-Beltian units parallels bedding, striking northeast-southwest

Table 1. Summary of mineral assemblages in the Kelly and Carter Creek areas of the Ruby Range

Mafic gneisses: Kelly assemblages = 1a-1c; Carter Cr. assemblages = 1d-1h.

- 1a) and (or 1ab) + sal ± hbl ± CaMg alm* ± qtz + ilm ± {apa, mag, zirc}**. Samples RMK-13,-17,-26,-27₁,-27₂, and -77.
 1b) and + CaMg alm + qtz + hyp + hbl ± bio + {sal} ± {mag, ilm, zirc}. Samples RMK-27₃ and -51.
 1c) lab + hbl + sal + hyp. Sample RMK-122.
 1d) mic + scap + hbl + sal + CaMn alm + sph + cal + {graph, qtz, zirc}. Samples RMC-1₁ and -1₂.
 1e) sal + FeMn gross + and + cal + sph + {apa}. Sample RMC-71.
 1f) hbl + and ± cum + CaMg alm + qtz ± bio ± ilm ± {apa}. Samples RMC-77,-97, and -123.
 1g) hbl + qtz ± and + MgCa (or CaMg) alm + cum ± ilm. Samples RMC-15D and -103A.
 1h) gedr + Mg alm + qtz + and + bio. Samples RMC-19E₁ and -19E₂.

Iron-formations: Kelly assemblages = 2a-2b; Carter Cr. assemblages = 2c-2d.

- 2a) qtz + mag + eul + {FeCa spess}. Sample RMK-48.
 2b) qtz + mag + Fe-aug (or aug) + Fe-hyp (or eul) + CaMn (or MnCa) alm ± {Mn-ilm, cal}. Samples RMK-37₁,-37₂, -46,-60,-76, and -80.
 2c) qtz + mag + hem ± Fe-hyp + cum ± hbl (or act) + {mic, cal, apa}. Sample RMC-44.
 2d) qtz + mag + hem ± diop (or hyp) + ab + Na trem + riebeckite ± andr + {mic, cal, apa}. Samples RMC-37 and -135.

Meta-pelites: Kelly and Carter Cr. assemblage.

- 3a) qtz ± mic perth + olig + MgMn (or Mg) alm + bio ± cord ± sill ± {ilm, rut, mag, zirc, apa, cal, mu (relict), and/or ky (relict)}. Samples RMK-66,-98, and -127; RMC-30,-84,-93,-112, and -151.

Meta-dolomites: Kelly assemblage.

- 4a) cal + dol + fo + sal + phl. Sample RMK-136.

In assemblages 1a-1b, coronas of alm-qtz symplectite on sal grains are common in plagioclase-rich horizons. In assemblages 2a-2b, incipient rimming of pyroxenes by grun and hbl is sometimes observed. In Sample RMK-98 (assemblage 3a), cord occurs as coronas on alm porphyroblasts. In assemblage 4a, fo is pseudomorphed by serp + mag. *Denotes alm garnet with appreciable gross (Ca) and lesser py (Mg) components. Similar notation is used elsewhere in the table. ** Brackets {} indicate accessory phases. Abbreviations: ab = albite; alm = almandine; and = andesine; andr = andradite; apa = apatite; aug = augite; bio = biotite; cal = calcite; cord = cordierite; cum = cummingtonite; diop = diopside; dol = dolomite; eul = eulite; fo = forsterite; gedr = gedrite; graph = graphite; gross = grossularite; grun = grunerite; hem = hematite; hbl = hornblende; hyp = hypersthene; ilm = ilmenite; ky = kyanite; lab = labradorite; mag = magnetite; mic = microcline; mu = muscovite; olig = oligoclase; perth = perthite; phl = phlogopite; qtz = quartz; riebeckite; rut = rutile; sal = salite; scap = scapolite; serp = serpentine; sill = sillimanite; sph = sphene; trem = tremolite; zirc = zircon.

Table 2. Metamorphic temperature estimates for the Kelly and Carter Creek areas of the Ruby Range

Mineral-pair	Temperature estimates (°C) (Kelly area)		Temperature estimates (°C) (Carter Creek area)		Reference
	Peak	Retrograde	Peak	Retrograde	
Cpx-opx	770±70 (9)	---	---	---	Wood and Banno (1973)
Cpx-opx	700±100 (9)	---	---	---	Ross and Huebner (1975)
Gar-opx	760±40 (3)	---	---	---	obtained by comparison with data from Dougan (1974); see text
Gar-cpx	745±45* (5)	690±35 (5)	665±60 (3)	---	Råheim and Green (1974); Mysen and Heier (1972)
Gar-hb	740±80 (4)	<740±80 (4)	645±70 (6)	<645±70 (1)	Saxena (1968); Turner (1968)
Gar-bi	760±50 (1)	635±50 (3)	665±50 (3)	590±50 (5)	Thompson (1976)
Ksp-plag**	735±70 (1)	---	660±40 (1)	---	Whitney and Stormer (1977)
Sill-ksp	>630 (4)	---	>630 (9)	---	minimum thermal stability at $a_{H_2O} = 0.25$, $P_I = 6$ kbar; calculated by Wintsch (1975)
Gar-cd	665±50* (1)	545±50 (1)	---	---	Thompson (1976)
Cal-dol	---	545±45 (1)	---	---	Sheppard and Schwarcz (1970)
Qtz-mag	---	540±15 (3)	---	490±20 (3)	Bottinga and Javoy (1973)
Qtz-mag	---	530±15 (3)	---	475±25 (3)	Becker and Clayton (1976)
Consensus peak temp.	745±50°C		675±45°C		

The value reported for each geothermometer represents either the temperature range indicated by the various samples analyzed, or the published uncertainty of the method (if known) applied to the average temperature obtained by that method. Number of samples analyzed is given in parentheses next to each temperature estimate. Abbreviations: *cpx* = clinopyroxene, *opx* = orthopyroxene, *gar* = garnet, *hb* = hornblende, *bi* = biotite, *ksp* = K feldspar, *plag* = plagioclase, *cd* = cordierite, *cal* = calcite, *dol* = dolomite, *qtz* = quartz, *mag* = magnetite, and *sill* = sillimanite.

*It cannot be stated with certainty that these estimates reflect peak conditions (see text for explanation); consequently, these estimates are not figured into the consensus peak temperature for the Kelly area. Gar-cpx estimates assume $P_I = 6.5$ - 6.8 kbar (Kelly) and 5.0 - 5.7 kbar (Carter Creek).

**Ksp-plag estimates assume $P_I = 8$ kbar (Kelly) and 6 kbar (Carter Creek).

and dipping to the northwest. Widespread alteration of prograde assemblages indicates that Precambrian greenschist- to lower amphibolite-facies metamorphism occurred in the Ruby Range. Whether this metamorphism occurred during the waning stages of the 2700 m.y. event or represents a separate event cannot be discerned from field or petrographic evidence. However, Giletti (1966) has reported a number of 1600 m.y. mineral dates for the area. Also, Reid and Greenwood (1968) have documented a 670 m.y. event in Idaho, which may have affected basement rocks in Montana as well. These dates indicate a polymetamorphic history for the area.

The "Cherry Creek" rocks of the Kelly and Carter Creek areas were mapped in detail by James *et al.* (1969) and James and Wier (1972a, b). Additional details on the stratigraphy, structure, and petrology of these rocks are given in Heinrich (1960), Dahl (1977), and Garihan (1979).

Analytical procedures

Electron microprobe procedures

Chemical analyses for nine oxide components were performed on coexisting minerals in some thirty

polished thin sections using the 3-spectrometer ETEC Autoprobe at Indiana University; operating conditions and analytical procedures were identical to those described in Klein (1974). Representative analyses of minerals—too numerous to list here—are catalogued in Dahl (1977). Averaged (mean) microprobe analyses (uncorrected for generally small amounts of Fe^{3+}) were used in the calculation of distribution coefficients which, in turn, were referred to the published mineral-pair geothermometers and grade indicators itemized in Table 2 and Figure 2.

Oxygen isotopic procedures

Quantitative oxygen extraction of quartz and magnetite from iron-formations was performed by the BrF_5 method described by Clayton and Mayeda (1963). Liberated oxygen was then passed over a heated graphite disc for quantitative conversion to CO_2 . Isotopic composition of the CO_2 was determined with a 15-cm, 60° -sector mass spectrometer equipped with a double inlet and a double collector similar to that described by McKinney *et al.* (1950). Oxygen yields of 100 ± 3 percent were obtained for all quartz runs and most of the magnetite runs. Some yields for magnetite runs were as low as 85 percent,

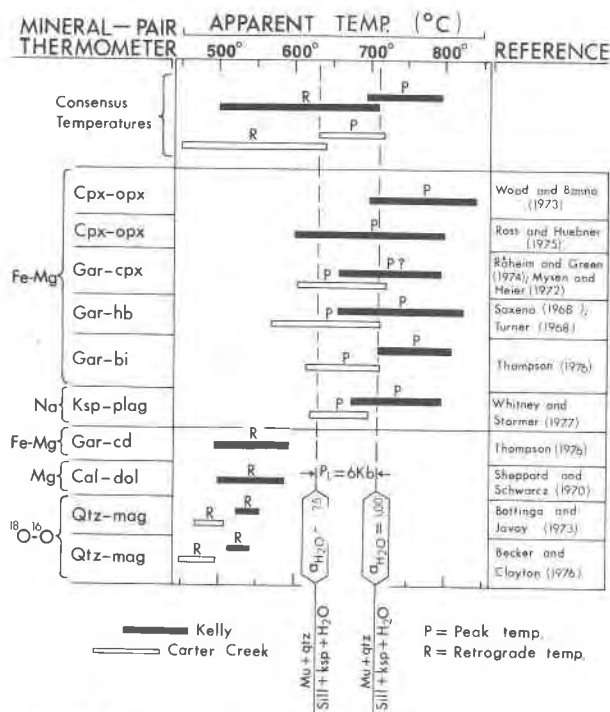


Fig. 2. Summary of temperature estimates obtained from various mineral-pair geothermometers for high-grade metamorphic rocks of the Ruby Range. For a given thermometer, the horizontal bar represents either the temperature range indicated by the various samples analyzed, or the published uncertainty of the method (if known) applied to the mean temperature obtained by that method. Question mark indicates uncertainty in the interpretation of K_D . Breakdown temperatures of muscovite + quartz at $P_1 = 6$ kbar for $a_{H_2O} \approx 0.25$ and $a_{H_2O} = 1.0$ estimated from Kerrick (1972) and Chatterjee and Johannes (1974), respectively. See Table 2 for mineralogical abbreviations.

but no detectable error was introduced in measured isotope ratios, as determined by comparison with replicate runs that gave 100 percent yields. Most runs were performed in replicate, and a 95 percent confidence limit of ± 0.12 per mil was recorded for $\delta^{18}O$ in most samples.

Geothermometric results and discussion

Introduction

Metamorphic temperature estimates based upon nine different mineral pairs are presented in Table 2 and Figure 2. These estimates are discussed individually in the following sections.

Clinopyroxene-orthopyroxene

Rim compositions of coexisting pyroxenes in mafic granulite gneisses and iron-formations from the Kelly area were referred to two widely-used pyrox-

ene geothermometers. The first of these methods—that of Wood and Banno (1973)—uses an equation based on Davis and Boyd's (1966) experimental determination of the diopside-enstatite solvus. Their equation contains an empirical correction for the FeO content of orthopyroxene and assumes ideal mixing of the $MgSiO_3$ component in diopside coexisting with orthopyroxene. The Wood-Banno method yields a mean temperature estimate of $770 \pm 70^\circ C$ for the Kelly area, based on nine pyroxene-bearing samples. The $\pm 70^\circ C$ uncertainty reflects the range of individual estimates; a $\pm 60^\circ C$ uncertainty is placed on the method by Wood and Banno. The more Mg-rich pyroxene pairs from the mafic granulite gneiss assemblages consistently give the higher temperatures in the above range, whereas pyroxenes from the iron-formations yield the lower temperatures. A similar trend was observed and explained by Wells (1977). Hewins (1975) concluded that the Wood-Banno geothermometer may read too high by as much as $50^\circ C$ for granulite-facies assemblages. Indeed, the temperature range calculated by this method is higher than that predicted by any other method (see Table 2 and Fig. 2).

Ross and Huebner (1975) performed heating experiments on natural, exsolved pyroxenes at various temperatures under controlled oxygen fugacity (iron-wustite buffer) at 1 atm. Electron microprobe analyses of these exsolved pyroxenes permitted them to contour the pyroxene solvus for temperature on a Wo-En-Fs diagram. Dahl (in preparation) showed that pyroxene pairs from the Kelly area plot along the $700^\circ C$ isotherm defined by Ross and Huebner [as do pyroxene pairs from the nearby Tobacco Root Mountains (Immega and Klein, 1976)]. Mori and Green (1976) suggest that the diopside-enstatite solvus is insensitive to variations in load pressure (P_1) below $900^\circ C$. If this relationship extends also to pyroxenes containing Fe, it follows that the $700^\circ C$ estimate need not be corrected for pressure. Ross and Huebner place a $\pm 100^\circ C$ uncertainty on estimates obtained by their method.

Garnet-orthopyroxene

The sub-assemblage garnet-orthopyroxene is common in the mafic granulite gneisses and iron-formations of the Kelly area, and neither mineral exhibits chemical zonation. Although a geothermometer based on this mineral pair has never been calibrated, Dahl (in preparation) found that the value of $K_D(Fe/Mg; \text{gar-opx})$ is dependent on temperature and on the Mn content in garnet (Mn^{Gar}). By plotting data

from various world-wide localities on a graph of K_D vs. Mn^{Gar} , it was possible to infer relative metamorphic temperatures among these areas from the relative positions of their respective isotherms. Because the isotherm for the Kelly area merged with an isotherm defined by garnet-orthopyroxene-bearing gneisses from the Venezuelan Guyana shield (Dougan, 1974), it was concluded that these two areas experienced similar peak metamorphic temperatures. Thus, Dougan's estimate of $760 \pm 40^\circ\text{C}$, which was partly based on garnet-cordierite geothermometry, is tentatively applied here as an indirect estimate for the Kelly area. This estimate agrees well with the consensus peak estimate of $745 \pm 50^\circ\text{C}$ for the Kelly area given in Table 2, although it must be stated that the effects of load pressure on isotherms in K_D - Mn^{Gar} space are not known. These effects are assumed to be small.

Garnet-clinopyroxene

This mineral pair, like garnet-orthopyroxene, is common in the mafic granulite gneisses and iron-formations of the Kelly area. It is found also in several calcite-bearing samples of mafic gneiss from the Carter Creek area.

Råheim and Green (1974) experimentally calibrated the distribution coefficient $K_D(\text{Fe}/\text{Mg}; \text{garnet-clinopyroxene})$ as a geothermometer for eclogites, using mineral mixes and artificial tholeiitic glasses as starting materials. Their piston-cylinder experiments yielded the temperature-dependence of K_D from 600° to 1400°C at 30 kbar, and the pressure-dependence of K_D from 20 to 40 kbar at 1100°C . Based on straight-line trends at these conditions, they extrapolated their results to other pressures and temperatures in the eclogite field and, tentatively, to conditions within the garnet granulite field. One difficulty involved in applying this geothermometer to garnet-clinopyroxene-bearing rocks of the Ruby Range is that these rocks formed at load pressures well below the 20–40 kbar range studied by Råheim and Green, as evidenced by the abundance of sillimanite (and not kyanite) in associated meta-pelites. A second difficulty is that a marked dependence of K_D on Mn^{Gar} exists among the low-Mn mafic granulite gneisses and the relatively high-Mn iron-formations of the Kelly area (Dahl, in preparation). Thus, to obtain a temperature estimate for the Kelly area, it is necessary to choose only K_D values that correspond to the range of Mn contained in Råheim and Green's experimentally-derived garnets. Accordingly, K_D values of 5.5 (core) and 6.6 (rim) were chosen; they yield respective temperature

estimates of 700° and 660°C (assuming $P_L = 8$ kbar) for the Kelly area.

A second garnet-clinopyroxene geothermometer was calibrated empirically by Mysen and Heier (1972), using published analyses of low-Mn garnet-clinopyroxene pairs together with the published equilibration temperature for each assemblage. As noted by Råheim and Green, this approach depends on the validity of the P_L - T estimates for the different metamorphic terrains, and most of these estimates carry large uncertainties. Secondly, the range of pressure estimates (6 to 36 kbar) is so large that, even if the pressure effect on K_D is small, data over so large a pressure range probably cause some of the broad scatter observed in Mysen and Heier's calibration. Thirdly, their calibration is based on chemical analyses of separated minerals, and the possibility of chemical zoning in these minerals adds uncertainty to estimated K_D values. To allow for these three sources of error, Mysen and Heier place a $\pm 60^\circ\text{C}$ uncertainty on temperature estimates obtained by their method. The Mn-normalized core and rim K_D values from the Kelly area, when referred to Mysen and Heier's geothermometer, yield respective estimates of $790 \pm 60^\circ$ and $725 \pm 60^\circ\text{C}$. These estimates are significantly higher than those obtained by Råheim and Green's method, which may in part reflect the high-pressure (mean ≈ 20 kbar) assemblages used in Mysen and Heier's calibration. Indeed, Råheim and Green's calibration of K_D as a function of P_L and T shows that, for load pressures of 21 and 8 kbar, a K_D value of 6.6 yields respective temperatures of 745° and 655°C . This means that for the Kelly area ($P_L \approx 8$ kbar), temperatures estimated by the Mysen and Heier geothermometer are probably somewhat high.

The petrologic significance of core and rim temperature estimates remains to be resolved. Clinopyroxene (salite) and almandine in the mafic granulite gneisses of the Kelly area generally appear to be in textural equilibrium, but coronas of garnet-quartz symplectite on clinopyroxene grains in contact with plagioclase are not uncommon, as shown in Figure 3. This texture, along with zoning profiles in andesine, suggests that the overall garnet-forming reaction:



was in progress. Manna and Sen (1974) studied similar coronal garnets in granulite gneisses from Saltora, India, and concluded that the garnets formed from clinopyroxene during isobaric cooling of gabbroic

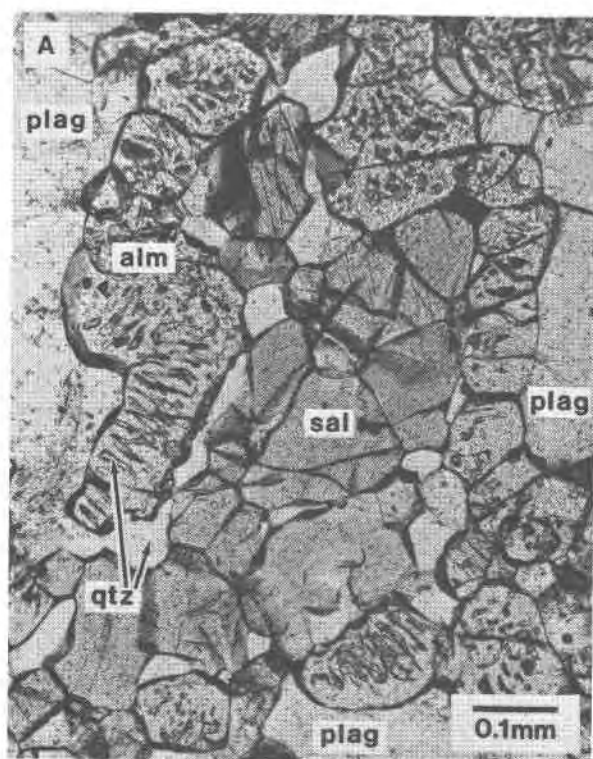


Fig. 3. Corona of almandine-quartz symplectite, separating salite from plagioclase (andesine) in mafic granulite gneiss (sample RMK-27₂) of the Kelly area. This texture suggests that almandine and quartz formed at the expense of salite and plagioclase. Plane polarized light.

magma at great depth. Thus, the 660° and 725°C estimates, which are based upon rim K_D values, may represent the temperature of last equilibration between garnet and its parent clinopyroxene during isobaric cooling. A mean temperature of $690 \pm 35^\circ\text{C}$ is reported in Table 2.

The petrologic significance of core compositions is ambiguous, because clinopyroxenes in the mafic granulite gneisses exhibit bell-shaped zoning profiles for the elements Na, Al, Fe, and Mg, whereas the garnets mantling these pyroxenes (see Fig. 3) are typically unzoned. The rims of these pyroxenes are depleted in Na, Al, and Fe and enriched in Mg relative to the cores (Dahl, 1977; Fig. 13). These profiles are consistent with the textural evidence that almandine in mafic gneisses formed at the expense of clinopyroxene. Prior to garnet growth, however, it is likely that the clinopyroxene was chemically homogeneous (as are pyroxenes in the iron-formations) and had a composition similar to the present core composition. It follows from the pyroxene zoning profiles described above that the first garnet grains to form would have equilibrated with clinopyroxenes whose

rims were less depleted in Fe and less enriched in Mg than the present rims (*i.e.*, the Fe/Mg ratio at the rims was once higher than that observed today). If this were the case at the time of first garnet growth, then the value of the rim $K_D(\text{Fe/Mg}; \text{gar-cpx})$ must have been lower than it is now, indicating a temperature higher than that of final equilibration. Thus, the garnet-clinopyroxene core and rim "temperatures" may give a qualitative indication as to the direction of temperature change during garnet growth from clinopyroxene, but it cannot be stated that the core "temperature" represents peak metamorphic conditions. The uncertainty concerning this estimate is noted in Table 2 and Figure 2.

It should be possible to derive a metamorphic temperature estimate for the Carter Creek area, based on local occurrences of coexisting garnet and clinopyroxene. Garnet-clinopyroxene pairs in mafic gneisses from the Carter Creek area differ, however, from their Kelly area counterparts in that (1) both minerals are chemically unzoned (except in sample RMC-71, in which a prograde zoning profile is preserved in the garnet), (2) they are never found in a coronal relationship, (3) both are much more calcic, and (4) they exhibit significantly higher K_D values (*ca.* 11.5; Dahl, in preparation). Dahl concluded that the higher K_D values for the Carter Creek samples resulted from a combination of lower metamorphic temperature and the much higher calcium content of the garnets (Ca^{Gar}). Thus, comparison of Kelly and Carter Creek garnet-clinopyroxene pairs of approximately equal Mn content, combined with knowledge of the effect of Ca^{Gar} on K_D for amphibolite-facies rocks (Banno, 1970), permits a tentative peak temperature estimate of 665°C for the Carter Creek area. This estimate is concordant with other estimates of peak temperature for the area, as shown in Table 2 and Figure 2.

In summary, the temperature estimates obtained from garnet-clinopyroxene pairs are consistent with the other estimates given in Table 2 and shown in Figure 2. However, rigorous application of the available gar-cpx geothermometers is difficult for the reasons outlined above.

Garnet-hornblende

Coexisting garnet and hornblende are common in mafic gneisses from both the Kelly and Carter Creek areas. Saxena (1968) showed that $K_D(\text{Fe})$ for this mineral pair is a function of metamorphic grade. According to his empirical calibration, based upon published compositional data, garnet-hornblende pairs

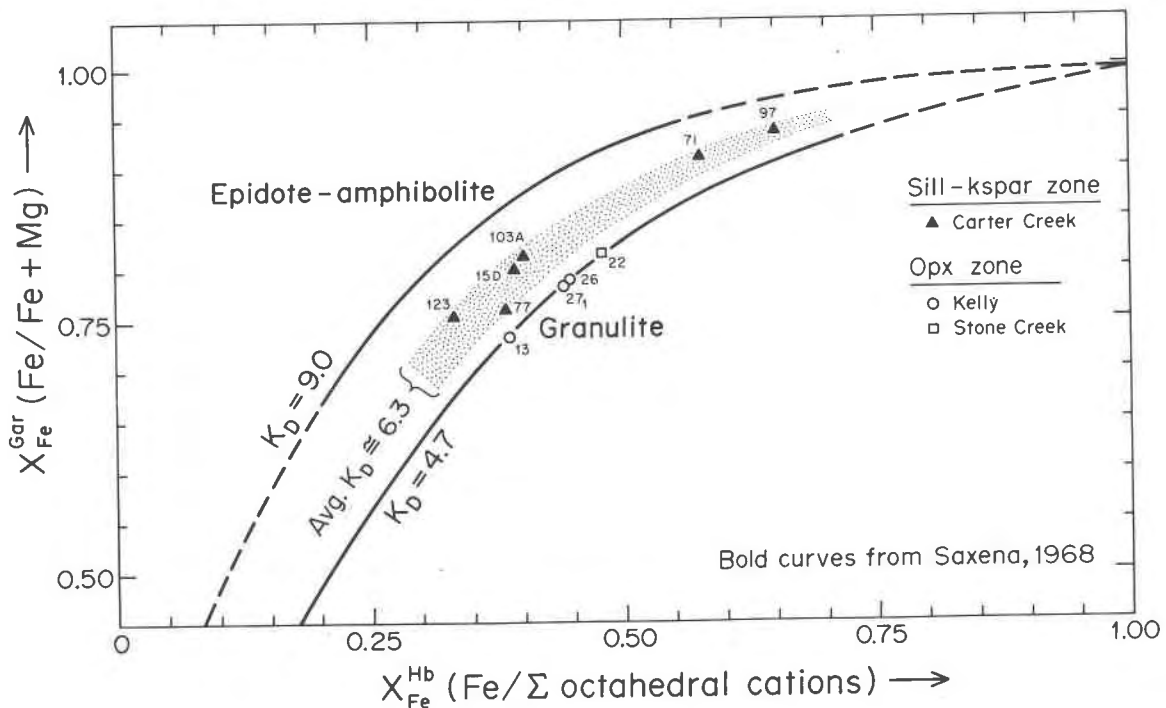


Fig. 4. Distribution of iron between coexisting garnet and hornblende in mafic gneisses of the Ruby Range. Stippled band defines the distribution for Carter Creek samples (mean $K_D = 6.3$), whereas the Kelly and Stone Creek samples plot along Saxena's (1968) trend for granulites (mean $K_D = 4.7$). Epidote-amphibolite trend (mean $K_D = 9.0$) is also from Saxena. $K_D(\text{Fe}; \text{gar-hb}) = [X_{\text{Fe}}^{\text{Gar}} / (1 - X_{\text{Fe}}^{\text{Gar}})] / [X_{\text{Fe}}^{\text{Hb}} / (1 - X_{\text{Fe}}^{\text{Hb}})]$. Except for sample RMC-71, all points represent core K_D values calculated from mean microprobe analyses of adjacent garnet and hornblende grains. The rim K_D value is plotted for sample RMC-71 because garnet in this sample exhibits prograde zonation.

in epidote-amphibolite-facies rocks have K_D values of about 9.0, whereas in granulite-facies rocks, K_D values of approximately 4.7 are typical. These calibration curves are shown in Figure 4, along with garnet-hornblende data from the Ruby Range. Three data points representing granulite-facies assemblages of the Kelly area plot directly on Saxena's curve for granulites. On the other hand, six data points representing the upper-amphibolite-facies assemblages of the Carter Creek area define a band that plots at somewhat lower grade than the granulite trend. Using P - T boundaries assigned by Turner (1968, p. 366) for the various metamorphic facies, the granulite-facies trend defined in Figure 4 for the Kelly area is assigned a broad peak temperature range of $740 \pm 80^\circ\text{C}$. Similarly, the upper-amphibolite-facies trend of the Carter Creek samples corresponds to a peak temperature range of $645 \pm 70^\circ\text{C}$. These temperature ranges are concordant with other estimates given in Table 2 and Figure 2.

The trends in Figure 4 are based on core K_D values. Rim compositions of both garnet and hornblende are slightly different from core compositions, and probably indicate incipient mass transfer of Fe

and Mg between the two minerals, in response to declining temperatures of metamorphism. This conclusion is supported by antithetic compositional profiles in the two minerals and by the fact that rim K_D values (not shown in Fig. 4) are invariably a little higher than the corresponding core values. Apparent retrograde (rim) temperatures are reported in Table 2.

Garnet-biotite

Coexisting garnet and biotite are common in pelitic rocks of the Kelly and Carter Creek areas. Values of $K_D(\text{Fe}/\text{Mg}; \text{gar-bi})$ for these rocks, and corresponding temperature estimates obtained from the theoretical geothermometer of Thompson (1976), are presented in Table 3 for both core and rim compositions. In most samples, temperatures based on core K_D values are higher than those based on rim values, and mass transfer of Fe and Mg at garnet-biotite contacts is inferred on the basis of antithetic compositional profiles determined by the electron microprobe. Thus, core compositions probably reflect peak temperatures, and rim compositions reflect temperatures that were set during retrograde cooling. This

Table 3. Metamorphic temperature estimates based on core and rim values of $K_D(\text{Fe/Mg}; \text{gar-bi})$

Sample	Rim		Core		Garnet zoning*	Average temp. ($^{\circ}\text{C}$)		
	K_D	$T(^{\circ}\text{C})$	K_D	$T(^{\circ}\text{C})$		Rim	Core	
Kelly	RMK-51 ₁	4.53	620		n.a.	635 \pm 50	760 \pm 50	
	RMK-51 ₂	4.31	640		n.a.			
	RMK-66	4.24	640	3.00	760			retrograde
	(RMK-127)	5.96	545)**					n.a.
Carter Creek	RMC-30	5.46	595	3.82	675	retrograde	590 \pm 50	665 \pm 50
	RMC-19E ₁	4.89	600			n.a.		
	RMC-19E ₂	4.57	620			n.a.		
	RMC-112	5.53	555	4.06	655	retrograde		
	RMC-151	5.18	585	3.87	670	retrograde		
	(RMC-93)	4.75	605	5.23	585)**	prograde		
	(RMC-84)	5.72	560	4.73	605)**	retrograde		

Temperatures are calculated from Thompson's (1976) calibration. Approximate uncertainties are $\pm 50^{\circ}\text{C}$.

* Prograde vs. retrograde zoning in garnet is figured from the criteria of Sturt (1962)-- n.a. = not analyzed.

** Data from these samples are not figured into the average temperature estimates, as explained in the text.

conclusion is supported by the criteria of Sturt (1962) for garnet zoning, namely, that garnets which contain more Cc+Mn and less Fe+Mg in their rims than in their cores exhibit retrograde zoning. The nature of garnet zoning in each sample is included in Table 3.

Thompson's geothermometer yields mean peak estimates of $760 \pm 50^{\circ}$ and $665 \pm 50^{\circ}\text{C}$ for the Kelly and Carter Creek areas, respectively ($\pm 50^{\circ}\text{C}$ is the uncertainty quoted by Thompson for the method). These estimates are concordant with peak estimates obtained by other methods, as indicated in Table 2. Samples RMK-127 and RMC-84 are not included in the averages because, in these samples, high Mn^{Grt} values appear to have contributed to high values of K_D , resulting in artificially low temperature estimates. The core K_D of sample RMC-93 is also ignored, because garnet in this sample exhibits a prograde zoning profile, making it impossible for the core K_D to represent peak metamorphic conditions.

K feldspar-plagioclase

Stormer (1975) calculated the P_L - T dependence of albite distribution between coexisting alkali feldspar and plagioclase, using published Margules parameters for alkali feldspar (sanidine) and assuming ideal mixing in plagioclase. Strictly speaking, this geothermometer is applicable only to sanidine-plagioclase pairs, because it has been shown experimentally (Müller, 1971) that the albite content of alkali feldspar coexisting with plagioclase is highly dependent on its structural state. Alkali feldspar in meta-pelites from the Ruby Range occurs primarily as microcline, and application of Stormer's geothermometer to mi-

crocline-plagioclase pairs in the Kelly and Carter Creek areas yields respective estimates of 675° (assuming $P_L \approx 8$ kbar) and 620°C (assuming $P_L \approx 6$ kbar). These temperatures, which are somewhat low compared to the consensus peak estimates given in Table 2 for the two areas, indicate that Stormer's sanidine-plagioclase geothermometer should not be applied to microcline-plagioclase pairs.

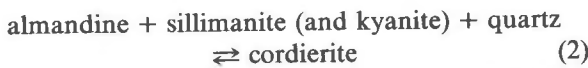
Whitney and Stormer (1977) recalibrated the two-feldspar geothermometer using Margules parameters for microcline (instead of for sanidine). This new method yields much higher estimates of 790° and 700°C for the Kelly and Carter Creek areas, respectively. According to Whitney and Stormer, however, microcline is not stable at temperatures much in excess of 500°C , and where it is found in high-grade metamorphic rocks, it probably transformed from feldspar that originally crystallized as orthoclase. Furthermore, they state that temperature values for feldspars crystallizing as orthoclase would probably lie between those defined for sanidine and microcline. Therefore, I suggest that alkali feldspars in meta-pelites of the Ruby Range originally crystallized in the range 675 – 790° (i.e., $735 \pm 60^{\circ}\text{C}$) and 620 – 700° (i.e., $660 \pm 40^{\circ}\text{C}$), respectively. These results are consistent with the consensus estimates reported in Table 2 and Figure 2.

Application of two-feldspar thermometry requires not only an accounting of feldspar structural state, but also consideration of an analytical difficulty, namely, chemical inhomogeneity of feldspars. Plagioclase in the Ruby Range meta-pelites usually exhibits weak normal zoning, and the coexisting *K* feldspar is commonly perthitic, with some parts of

the grains slightly more perthitic than others. It is not clear, therefore, which feldspar compositions were in equilibrium at peak metamorphic conditions. The temperatures reported above were calculated on the basis of bulk K feldspar compositions (determined by the mean of repeated cross-sectional analyses of perthitic grains) and plagioclase core compositions. Such a procedure yields maximum estimates of temperature.

Garnet-cordierite

Four garnet-cordierite pairs in one Mg-rich metapelite (sample RMK-98) from the Kelly area yield rim K_D (Fe/Mg; gar-cd) values of 11.6. This K_D , when applied to the theoretical garnet-cordierite geothermometer of Thompson (1976), results in a final equilibration temperature of $545 \pm 50^\circ\text{C}$ (uncertainty quoted by Thompson), which is significantly lower than the consensus peak estimate of $745 \pm 50^\circ\text{C}$. In thin section, undeformed cordierite grains containing sillimanite and kyanite fragments are found as coronas on moderately-fractured, Mg-rich almandine porphyroblasts. This texture suggests that cordierite formed in the continuous reaction:



Rims of almandine grains in sample RMK-98 are markedly depleted in Mg and enriched in Fe relative to the cores. This zoning is probably the result of chemical potential gradients induced by the crystallization of coronal cordierite.

According to the experimental work of Hensen and Green (1973), the above reaction is highly sensitive to changes in load pressure, with decreasing load pressure favoring the growth of cordierite. Furthermore, Hollister (1977) suggested that almandine-cornal cordierite textures may signify uplift events. For the Ruby Range, therefore, it is tentatively concluded that coronal cordierite formed from almandine in response to a decrease in load pressure, which occurred during Precambrian uplift and erosion following the high-grade metamorphism. The 545°C garnet-cordierite temperature, then, must reflect thermal conditions at some time during the uplift event. The exact timing of this event is not clear; it could have occurred during the waning stages of the 2700 m.y. metamorphism, or during the subsequent greenschist-lower amphibolite-facies metamorphism (1600 m.y.?, 670 m.y.?).

Because reaction 2 has proceeded to the extent that only small fragments of almandine remain, it is not

likely that the interiors reflect the composition of almandine at peak metamorphic conditions. Therefore, temperature estimates based on these compositions are of doubtful significance and are not figured into the consensus peak estimate for the Kelly area.

Calcite-dolomite

Sheppard and Schwarcz (1970) used the experimental carbonate solvus data of Goldsmith and Heard (1961) and Goldsmith and Newton (1969) to derive a least-squares expression for the temperature-dependence of Mg distribution between coexisting calcite and dolomite at $400^\circ\text{C} < T < 1075^\circ$ and $P_L < 2\text{kbar}$. For $P_L \approx 10\text{kbar}$, the calculated solvus temperature is too high by about 25°C (Goldsmith and Newton, 1969). Variable amounts of Fe^{2+} , Mn^{2+} , and Sr^{2+} in calcites affect temperature estimates, but not at concentrations less than 1–2 mole percent (Sheppard and Schwarcz, 1970). For chemically unzoned, low Fe-Mn-Sr calcite-dolomite pairs in metadolomite sample RMK-136 (Kelly area), six pair analyses yield temperatures in the range $545 \pm 45^\circ\text{C}$. This estimate is far below the consensus peak estimate of $745 \pm 50^\circ\text{C}$, but is concordant with the garnet-cordierite temperature obtained from sample RMK-98. Carbonates are highly susceptible to recrystallization. Thus, it is possible that calcite and dolomite in sample RMK-136 underwent recrystallization (perhaps in response to a deformational pulse) during retrograde or subsequent low-grade metamorphism. Similar reasoning may explain a low calcite-dolomite temperature of 590°C reported by Sheppard and Schwarcz for a granulite-facies metadolomite from the Grenville Province. However, these low temperature estimates may reflect inaccuracy in the original calibration as applied to high-grade rocks.

Quartz-magnetite

Oxygen isotopic compositions ($\delta^{18}\text{O}$) and calculated fractionations (Δ_{Q-M}) for coexisting quartz and magnetite are presented in Table 4 for six iron-formation samples from the Ruby Range. The calculated values of Δ_{Q-M} for the Kelly samples are significantly less than those for the Carter Creek samples. The temperature-dependence of Δ_{Q-M} was calibrated by Becker and Clayton (1976), using a combination of calculation and experimental mineral-water equilibrium data. Isotopic temperatures based upon their calibration fall in the ranges $530 \pm 15^\circ$ (Kelly) and $475 \pm 25^\circ\text{C}$ (Carter Creek). These estimates are in good agreement with those calculated by the method of Bottinga and Javoy (1973) (see Table 2 and Fig.

Table 4. Oxygen isotopic compositions, fractionations, and temperatures for coexisting quartz and magnetite in iron-formations from the Ruby Range

Locality	Sample	$\delta^{18}\text{O}_{\text{qtz}}$ (‰)	$\delta^{18}\text{O}_{\text{mag}}$ (‰)	$\Delta_{\text{Q-M}}$ $1000 \ln \alpha_{\text{Q-M}}$	Isotopic temp. (°C)	Isotopic temp. range (°C)
Kelly	RMK-46	12.85 ± 0.12 (2)	4.35 ± 0.12 (2)	8.43 ± 0.17	528 ± 10	530 ± 15
	RMK-60	11.05 ± 0.12 (2)	2.77 ± 0.12 (2)	8.23 ± 0.17	537 ± 10	
	RMK-76	11.50 ± 0.12 (2)	2.84 ± 0.12 (2)	8.60 ± 0.17	520 ± 10	
Carter Creek	RMC-37	17.74 ± 0.12 (2)	7.77 ± 0.12 (2)	9.85 ± 0.17	460 ± 10	475 ± 25
	RMC-44	18.24 ± 0.12 (2)	8.94 ± 0.12 (2)	9.18 ± 0.17	495 ± 10	
	RMC-135	18.99 ± 0.17 (1)	8.86 ± 0.12 (2)	9.99 ± 0.21	455 ± 10	

Temperatures obtained from the calibration of Becker and Clayton (1976). Uncertainties listed reflect only those inherent in the corresponding fractionations ($\Delta_{\text{Q-M}}$). 95% confidence limits and number of runs (in parentheses) are indicated for each $\delta^{18}\text{O}$ value. Low temperatures for samples RMC-37 and -135 may be due in part to fine magnetite grain sizes (mean 0.1 mm), compared to 0.6 mm for the other four samples.

2). Clearly, however, these temperatures must be regarded as poor indicators of peak metamorphic conditions, since they fall some 200°C below all other independent estimates. Mineralogical evidence is consistent with this conclusion. Specifically, meta-pelites associated with the iron-formations indicate that P_L - T conditions of at least sillimanite-K feldspar-zone metamorphism were attained throughout the Ruby Range. The minimum thermal stability of sillimanite + K feldspar at $P_L = 6$ kbar is roughly 700–715°C at $a_{\text{H}_2\text{O}} = 1$ (Chatterjee and Johannes, 1974; Kerrick, 1972) and 630°C at $a_{\text{H}_2\text{O}} \approx 0.25$ (Wintsch, 1975). Thus, even if $a_{\text{H}_2\text{O}}$ were as low as 0.25, the minimum peak temperature in the Ruby Range was approximately 630°C—far above the isotopic temperature. Furthermore, the common occurrences of the sub-assembly garnet±clinopyroxene±orthopyroxene in iron-formations and mafic gneisses of the Kelly area are clear indicators of granulite-facies metamorphism.

The low isotopic temperatures are probably the result of (1) solid-state diffusion of oxygen isotopes between quartz and magnetite during retrograde metamorphism, or (2) recrystallization of quartz and magnetite during retrograde or subsequent low-grade metamorphism. Alternatively, the low estimates point to errors at the high-temperature ends of published quartz-magnetite calibration curves. In the following discussion, however, it is assumed that these calibration curves are accurate and applicable to the Ruby Range iron-formations.

Let us first consider model (1), illustrated in Figure

5 for the Kelly area. During the sedimentary and diagenetic stages, equilibrium between quartz and magnetite was presumably maintained through mineral-water isotopic exchange with a common pore fluid. Continuous mineral-water isotopic re-equilibration, favored by fine quartz-magnetite grain sizes, probably took place in response to gradually increasing metamorphic temperatures. By the peak of metamorphism (745 ± 50°C in the Kelly area), quartz and magnetite were recrystallized into coarser grains (mean diameter 0.6 mm), and water was effectively removed by (1) the completion of dehydration reactions and (2) the diminution of pore space brought on by compaction and deformation. Temperatures were now probably high enough, however, for quartz-magnetite equilibrium to be maintained directly by solid-state diffusion of isotopes. This process probably maintained the equilibrium even as temperatures began to decline, at first keeping up with the change in temperature. But as temperatures decreased further, the kinetics of solid-state diffusion must have become increasingly less favorable, such that diffusion of isotopes slowed and eventually ceased, “freezing in” an apparent isotopic “temperature” of 530 ± 15°C in the Kelly area. Actual temperature continued to decline as the metamorphic event drew to a close.

Evidence in support of model (1) is provided by experimental diffusion data. Choudhury *et al.* (1965) studied the diffusion of oxygen in quartz at 667°C and determined diffusion coefficients (D) of $10^{-13.1}$ cm²/sec (perpendicular to the c axis) and

$10^{-11.4}$ cm²/sec (parallel to the *c* axis). If, as suggested by Clayton *et al.* (1968), quartz is among the most resistant of all minerals to isotopic exchange, then it follows that diffusion of oxygen in quartz is the rate-limiting factor in isotopic exchange with a coexisting mineral, such as magnetite. Thus, we can estimate the minimum exchange rate (*i.e.*, the maximum transfer time) of oxygen between coexisting quartz and magnetite in terms of quartz diffusion data alone. If quartz and magnetite grains (0.6 mm mean diameter) were in isotopic equilibrium at 667°C, and temperature were then to drop 10°C very slowly, the maximum time required to effect isotopic re-equilibration would be roughly equal to the time required for oxygen to diffuse from the core of a quartz grain to the rim of an adjacent magnetite grain. This time, calculated from known grain sizes and the larger *D* value given above, is approximately 1400 years. Thus, if the retrograde cooling rate were no greater than 1°C per 140 years over the assumed 10°C interval, isotopic re-equilibration could keep pace with the temperature decrease. However, steadily decreasing diffusion rates over a larger temperature span would cause the diffusion to lag behind the temperature decrease and eventually cease entirely, as shown schematically in Figure 5. In the absence of diffusion data for quartz at temperatures other than 667°C, a more quantitative diffusion model cannot be presented. However, the above calculations indicate that in a regionally-metamorphosed terrain, retrograde cooling over hundreds of thousands of years could result in significant oxygen isotopic re-equilibration between quartz and magnetite.

In terms of model (1), the significantly higher temperatures preserved by ferromagnesian silicate pairs (Fig. 2) indicate that, during retrograde cooling, solid-state diffusion of Fe and Mg between these silicates was much more sluggish than oxygen diffusion between quartz and magnetite. This relationship is shown in Figure 5 by the dashed lines labelled "GAR-CPX-OPX" and "QTZ-MAG." If the model is correct, then it follows that *D* values for Fe and Mg in garnet, pyroxene, hornblende, and biotite at a given temperature are much lower than those for oxygen in quartz and magnetite. The test for this prediction awaits the experimental determination of Fe-Mg diffusion coefficients for the ferromagnesian silicates at various temperatures.

An alternative model to explain the low oxygen isotopic temperatures is illustrated in Figure 5 [model (2)]. According to this model, isotopic temperatures in the two areas may have been set originally near

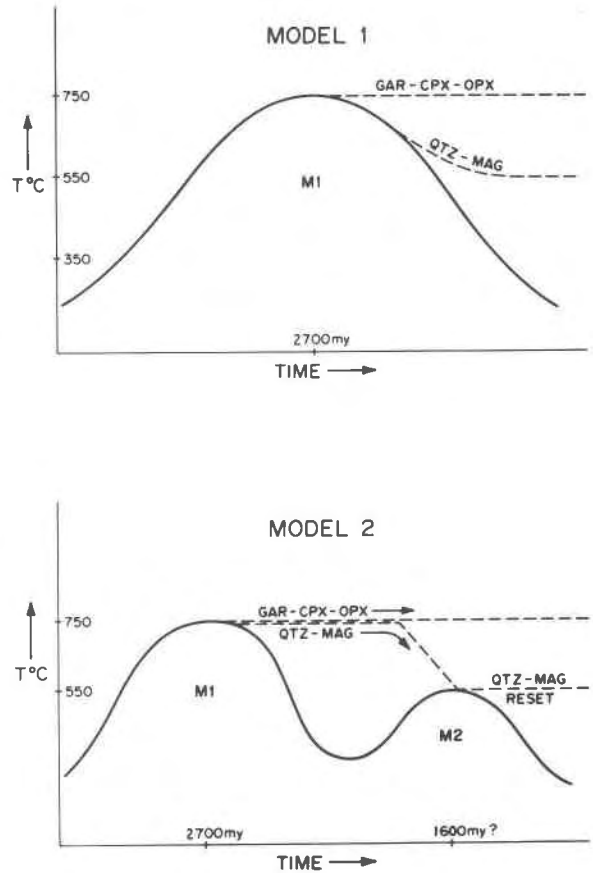


Fig. 5. Schematic metamorphic models (for the Kelly area) which possibly explain the divergence of oxygen isotopic temperatures from temperatures based on Fe-Mg distributions in coexisting ferromagnesian silicates. Dashed line indicates the presumed "temperature" that would be indicated by the given geothermometer at any time after the peak of metamorphism.

the peak of the high-grade metamorphic event (M1), but were subsequently reset by recrystallization of quartz and magnetite under retrograde conditions. Recrystallization may have occurred in response to a deformational pulse in the waning stages of the high-grade event, resulting in a higher retrograde temperature for the Kelly area than for the Carter Creek area. Alternatively, recrystallization in the two areas occurred during a subsequent low-grade event (M2), as discussed in earlier sections. Garnet, pyroxene, and other refractory silicates in the Kelly assemblages were perhaps more resistant to recrystallization than quartz and magnetite, and for this reason, they persisted as metastable relicts that record the $745 \pm 50^\circ\text{C}$ peak temperature of the high-grade event. This selective recrystallization model is shown schematically in Figure 5 by the diverging

dashed lines labelled "GAR-CPX-OPX" and "QTZ-MAG."

Support for model (2) comes from Wilson *et al.* (1970), who estimated peak metamorphic temperatures of 780°C for pyroxene granulites from the Musgrave Ranges, Central Australia. Post-metamorphic intrusion of the Ernabella Adamellite caused recrystallization of plagioclase, quartz, and magnetite in these granulites, resulting in quartz-magnetite and plagioclase-magnetite isotopic "temperatures" of around 550°C. Preservation of pyroxene at this temperature, however, suggests that this mineral did not recrystallize; dissolution of pyroxene at 550°C would most likely have culminated in amphibole crystallization.

Regardless of which of the two models in Figure 5 is applicable, it is noteworthy that re-equilibration of oxygen isotopes in the anhydrous Kelly iron-formation assemblages occurred despite the apparent lack of a fluid phase during retrograde metamorphism. Higher P_{H_2O} in the Carter Creek iron-formation (evidenced by a much higher modal amphibole/pyroxene ratio) may have facilitated oxygen isotopic exchange down to lower retrograde temperatures than in the Kelly iron-formation. In addition, the lower temperatures may be due in part to finer magnetite grain sizes. In samples RMC-37 and RMC-135, the mean magnetite grain size is 0.1 mm, compared to 0.6 mm in sample RMC-44 and in the three Kelly samples. Retrograde effects in very fine-grained magnetites were also reported by Shieh and Taylor (1969). Quartz grain sizes in the six samples studied are relatively uniform, ranging from 0.5 to 1.0 mm.

Conclusions

Mineral-pair geothermometry (based on six different pairs) and the coexistence of sillimanite and K feldspar in meta-pelites indicate peak metamorphic temperatures of $745 \pm 50^\circ$ and $675 \pm 45^\circ\text{C}$ for the Kelly and Carter Creek areas, respectively. These temperatures are consistent with the observed occurrences of orthopyroxene-zone and sillimanite-K feldspar-zone assemblages in the two areas. Furthermore, each mineral pair common to the two areas appears sensitive to the small metamorphic temperature difference inferred from mineral assemblages. The $\pm 45\text{--}50^\circ\text{C}$ uncertainty in temperature is determined from Figure 2 by defining the widest temperature range consistent with all the individual ranges shown for a given area. This uncertainty may help to define the limits of precision attainable for geothermometry in other high-grade terrains.

Concordant estimates of peak temperature are expected within a small area if (1) all geothermometers are accurately calibrated and applicable to the rocks in question, (2) all methods do in fact record peak conditions, (3) compositional and pressure effects on major-element and isotope distributions are negligible (or, if not, are taken into account), and (4) chemical equilibrium was approached during the exchange of species between coexisting minerals. The discordant estimates reported in Table 1 and Figure 2 indicate violation of one or more of these conditions. Orderly isotope distributions between quartz and magnetite (Table 4) suggest a close approach to chemical equilibrium [condition (4)]. Similarly, orderly element distributions between the rims of coexisting garnets and cordierites (in sample RMK-98) and between calcites and dolomites (in sample RMK-136) suggest attainment of equilibrium on a very local scale. Condition (3) is approximately satisfied, judging from the known chemistries of the mineral pairs and the known pressure effects on fractionation factors (Clayton *et al.*, 1972) and distribution coefficients. With regard to condition (1), there is no guarantee that the geothermometers giving discordant results in Figure 2 are accurately calibrated for high-temperature applications. If the published calibrations are accepted at face value, however, then the bimodal discordance shown in Figure 2 must indicate that different geothermometers record different "temperatures" [*i.e.*, condition (2) is violated].

Indeed, the interpretation of oxygen isotopic "temperatures," for example, has long been a controversial subject. Some investigators (*e.g.*, Taylor and Epstein, 1962, 1963; Taylor *et al.*, 1963) have maintained that these temperatures represent nearly peak thermal conditions. Other workers (*e.g.*, James and Clayton, 1962; Perry and Bonnicksen, 1966; Anderson *et al.*, 1971; Swulius, 1976) have concluded that minerals re-equilibrate by exchange of oxygen isotopes during retrograde cooling, thereby "freezing in" temperatures that may be significantly lower than peak temperatures. Deines (1975) has stated that the thermal significance of oxygen isotopic temperatures remains to be resolved.

Isotopic data from the Ruby Range iron-formations support the re-equilibration hypotheses. Specifically, it is thought that resetting of quartz-magnetite temperatures occurred in response to (1) solid-state diffusion of oxygen between quartz and magnetite during retrograde metamorphism or (2) recrystallization of the two minerals during retrograde or subsequent low-grade metamorphism. Garnet, py-

roxene, hornblende, biotite, and feldspars were perhaps more resistant to diffusion of chemical species and/or to recrystallization, and therefore retained compositions indicative of near-peak thermal conditions. Recrystallization at low temperature may also explain the low calcite-dolomite temperature reported for the Kelly area.

The temperature of final equilibration recorded for garnet-cordierite pairs ($545 \pm 50^\circ\text{C}$) is some 200° lower than the consensus peak estimate for the Kelly area. Because Thompson's (1976) geothermometer has been successfully applied in numerous high-grade terrains, however, the low estimate for the Kelly area is not thought to reflect major inaccuracies in his calibration. Instead, textural and chemical evidence suggests that cordierite in the Kelly area meta-pelites formed during retrograde metamorphism. The continuous reaction by which cordierite in the meta-pelites formed from garnet [reaction (2)] is favored by a decrease in load pressure, and could indicate a Precambrian uplift event post-dating the high-grade metamorphism. The concordant temperatures obtained from calcite-dolomite, quartz-magnetite, and garnet-cordierite pairs (*ca.* $545 \pm 50^\circ\text{C}$) may represent thermal conditions at some time during this event.

The results reported in this paper may help to resolve the petrologic significance of geothermometric data for other high-grade terrains. However, generalities regarding the exchange of chemical species during metamorphic events may not be entirely valid, inasmuch as different fluid partial pressures, different cooling rates, and different superimpositions of thermal and deformational maxima combine to make each metamorphic terrain unique.

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