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A revision of the garnet-clinopyroxene Fe^{2+}-Mg exchange geothermometer

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Abstract. A comprehensive experimental dataset was used to analyse the compositional dependence of the garnet-clinopyroxene Fe^{2+}/Mg partition coefficient (K_d). The Mg no. of garnet was found to have a significant effect on the K_d, in addition to calcium content of garnet. An empirical model was developed to relate these effects with equilibrium temperature and pressure in the form of a conventional geothermometer,

$$T(\mathrm{K}) = \{-1629[X^{\mathrm{Gt}}_{\mathrm{Ca}}]^2 + 3648.55[X^{\mathrm{Gt}}_{\mathrm{Ca}}] - 6.59[\mathrm{Mg\ no.\ (Gt)}] + 1987.98 + 17.66P\,(\mathrm{kbar})\}/(\ln K_d + 1.076).$$

Application of this thermometer produced reasonable temperature estimates for rocks from the lower crust (garnet amphibolites, granulites and eclogites) and the upper mantle (eclogite and lherzolite xenoliths in kimberlites, mineral inclusions in diamonds).

Introduction

Several experimental studies have demonstrated that Fe^{2+}/Mg partitioning between garnet and clinopyroxene is sensitive to external physical conditions and is strongly affected by internal compositional variations (e.g. Råheim and Green 1974a; Ellis and Green 1979; Pattison and Newton 1989). Råheim and Green (1974a) first calibrated the relationship between $K_d[= (Fe^{2+}/Mg)^{Gt}/(Fe^{2+}/Mg)^{Cpx}]$ and equilibrium temperature (T) and pressure (P) to form a geothermometer. Ellis and Green (1979) recognised a significant effect of Ca-content on K_d and quantified this effect to produce an improved version of the thermometer, which found a broad application in the field of petrology. Krogh (1988) further improved the thermometer by replacing the simple rectilinear relationship of $\ln K_d$ with $X^{Gt}_{Ca}[= Ca/(Ca + Mg + Fe^{2+} + Mn)]$ established by Ellis and Green (1979) with a curvilinear relationship. This latter improvement is confirmed by Brey and Köhler (1990), but the improvement is limited by the amount of experimental data used and by the exact mathematical model chosen to formulate the thermometer.

A recent experimental study by Pattison and Newton (1989) showed that the Mg no. $[= 100Mg/(Mg + Fe^{2+})]$ of garnet also had an important effect on K_d. They fitted their experimental data to a third order polynomial equation to account for this effect, and derived a new version of the thermometer that is applicable to rocks with Mg no. (Gt) ranging from 12.5 to 60. However, when applied to same mineral assemblages, this and other forms of the garnet-clinopyroxene Fe^{2+}/Mg exchange thermometer (e.g. Ellis and Green 1979; Ganguly 1979; Powell 1985) gave different temperature estimates.

Green and Adam (1991) specifically designed a new set of experiments to test various versions of this thermometer and found that none of the existing formulations could be safely applied to rocks of wide origin in terms of P, T and composition. In particular, Green and Adam (1991) found that the popular Ellis-Green thermometer reproduced their run temperatures for experiments at 30 kbar and 1150–1200 °C but increasingly overestimated equilibrium temperatures at lower pressures and lower temperatures. They also found that the Pattison-Newton graphic thermometer consistently underestimated their experimental temperatures but the Pattison-Newton mathematical formulation overestimated their experimental temperatures with two exceptions.

It is clear that additional experimental data are needed for further evaluation and improvement of this useful geothermometer. In this paper, I present new experimental results and use these in conjunction with published data to develop a refined empirical model to describe the partitioning behaviour of Fe^{2+}/Mg between garnet and clinopyroxene.

The dataset

The dataset used to analyse the Fe^{2+}/Mg partition between garnet and clinopyroxene consists of mineral equilibrium data in fifteen publications (Table 1) and new experimental data in ultramafic systems from Ai (1992, Table 2). The majority of these studies in the dataset are made on ultramafic compositions. Five series of experiments specifically address garnet-clinopyroxene equilibria. Two series of experiments are in mafic systems. They cover a wide range of "bulk rock" compositions. The experimental conditions range

Table 1. Reference sources for the experimental dataset

Reference no.	Source	Type of experiment	P (kbar)	T °C	No. of gt-cpx pairs
1	Råheim and Green (1974a)	Gt-cpx equilibria	20–40	600–1500	39
2	Råheim and Green (1974b)	Gt-cpx equilibria	12–30	1100–1300	6
3	Wood (1976)	Gt-cpx equilibria	20–45	1100–1400	8
4	Ellis and Green (1979)	Gt-cpx equilibria	15–30	750–1300	45
5	Pattison and Newton (1989)	Gt-cpx equilibria	15–29	700–1200	123
6	Johnston (1986)	Mafic system	23–30	1275–1455	7
7	Green and Adam (1991)	Mafic system	10–30	950–1200	17
8	Hensen (1973)	Ultramafic system	22.5–40.5	1110–1410	3
9	Akella and Boyd (1973)	Ultramafic system	30–45	1100	5
10	Akella (1976)	Ultramafic system	31–44	1100–1300	3
11	Mori and Green (1978)	Ultramafic system	30–40	950–1300	20
12	Nickel (1983)	Ultramafic system	35	1000–1400	10
13	Mengel and Green (1989)	Ultramafic system	25–28	975–1250	11
14	Brey et al. (1990)	Ultramafic system	28–60	900–1400	21
15	Wallace and Green (1991)	Ultramafic system	20–30	950–1100	15
16	Table 2	Ultramafic system	25–38	1200–1500	47

Gt, garnet; cpx, clinopyroxene

from 10 to 60 kbar in pressure and from 600 to 1500 °C in temperature, encompassing almost the entire lower crust-upper mantle region where garnet and clinopyroxene-bearing rocks originated. The dataset is therefore reasonably comprehensive and is meaningful for statistical analysis. However, the amount of data at low temperatures (below 900 °C) and high temperatures (above 1300 °C) is much less than in the middle temperature range (1000–1200 °C). The quality of these data is also not as good as that in the middle temperature range due to kinetic problems. At low temperatures, equilibrium is very difficult to achieve under laboratory conditions, especially in anhydrous systems. At high temperatures, the nucleation rate is so fast that metastable mineral compositions can often be preserved. Selection criteria such as cation totals to within ± 0.02 and weight percent of oxides to within ± 2.0 are employed to discriminate all experimental data available. The number of data is reduced to total 380 pairs.

Data analysis

Krogh (1988) proposed a general mathematical equation that governed the partitioning behaviour of Fe^{2+}/Mg between garnet and clinopyroxene:

$$\ln K_d = \frac{\{a[X_{Ca}^{Gt}]^2 + b[X_{Ca}^{Gt}] + c + dP(\text{kbar})\}}{T(\text{K})} + e \quad (1)$$

where a, b, c, d and e are constants. This agrees in principle with the forms of all previous derivations (Råheim and Green 1974a; Ellis and Green 1979; Ganguly 1979; Saxena 1979; Wells 1979; Dahl 1980; Powell 1985), except that of Pattison and Newton (1989) where the $\ln K_d$-pressure relationship is a kind of reciprocal relationship and is in apparent contradiction with the experimental observation that K_d increases with increasing pressure (Råheim and Green 1974a; Green and Adam 1991). Nonetheless, the experimental results of Pattison and Newton (1989) are important because they showed a clear dependence of K_d on Mg no. (Gt), particularly in the low Mg no. (Gt) region.

Figure 1 further demonstrates the negative correlation between $\ln K_d$ and Mg no. (Gt) (only experimental data at 30 kbar are shown for clarity). Although the restricted Mg no. range and scatter of data points obscured any correlation trend for data at 900 °C and 1500 °C, the trend is very obvious for data 1000–1400 °C. The effect of aluminium cations in clinopyroxene on K_d is insignificant, and is perhaps more dependent on bulk rock composition than temperature and pressure. A correlation between $\ln K_d$ and $X_{Ca}^{Cpx}[= Ca/(1 - Na)]$ (Ca and Na are cations in clinopyroxene for a 6-oxygen formulation) seems to be evident from Fig. 2, but this simply reflects close correlations of temperature with both $\ln K_d$ and X_{Ca}^{Cpx} (Bertrand and Mercier 1985; Brey and Köhler 1990). Although Ellis and Green (1979) implied that correction of the X_{Ca}^{Cpx} effect on K_d was accomplished by an X_{Ca}^{Gt} term, there does not seem to be a simple relationship between X_{Ca}^{Gt} and X_{Ca}^{Cpx} (Fig. 3). Therefore the dependence of K_d on X_{Ca}^{Cpx} is effectively taken into account by the temperature term and it follows that the composition terms which should be included in modelling the $\ln K_d$–T relation are X_{Ca}^{Gt} and Mg no. (Gt). Presence of Fe^{3+} in both garnet and clinopyroxene poses an additional problem for interpreting natural mineral assemblages (Luth et al. 1990), but at present this could not be evaluated with experimental data obtained from microprobe analysis.

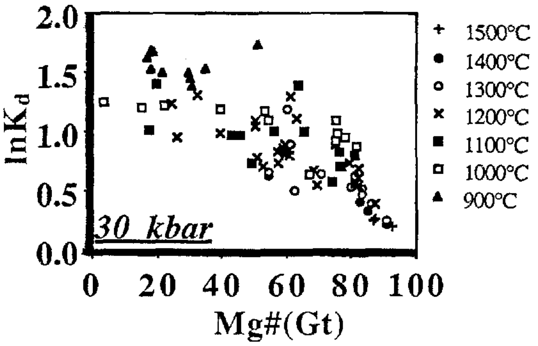

Fig. 1. The effect of Mg no. garnet, on $\ln K_d$. The correlation trends for most data at constant pressure and temperature are negative. For reasons of clarity, only data at 30 kbar are shown in this diagram (and Figs. 2, 3)

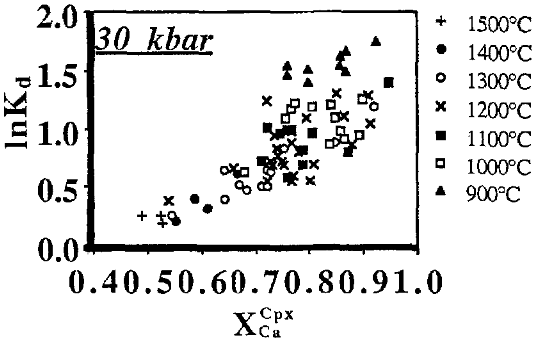

Fig. 2. The $\ln K_d - X_{Ca}^{Cpx}$ relationship. The parameter X_{Ca}^{Cpx} essentially reflects temperature. The three parameters, $\ln K_d$, X_{Ca}^{Cpx} and the reciprocal of temperature, are positively correlated. Only 30 kbar data are shown

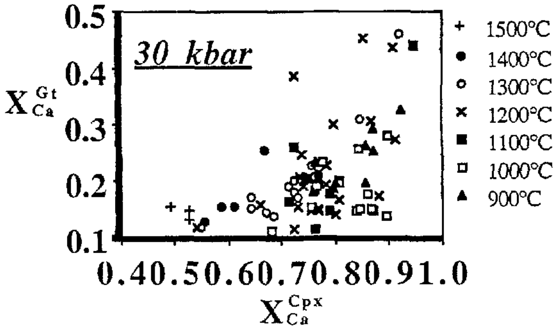

Fig. 3. The $X_{Ca}^{Gt} - X_{Ca}^{Cpx}$ correlation diagram. There is no clear, straight-forward relationship between X_{Ca}^{Gt} and X_{Ca}^{Cpx}. Only 30 kbar data are shown

Following Krogh (1988) and adding a Mg no. (Gt) term in addition to X_{Ca}^{Gt}, I derived a modified model to describe the partitioning behaviour of Fe^{2+}/Mg between garnet and clinopyroxene:

$$\ln K_d = \frac{\{a[X_{Ca}^{Gt}]^2 + b[X_{Ca}^{Gt}] + c[\text{Mg no. (Gt)}] + d + eP(\text{kbar})\}}{T(\text{K})} + f \qquad (2)$$

where a, b, c, d, e and f are constants. A statistical analysis of the 380 pairs of data in Tables 1 and 2 suggests that in a normal probability plot, data points which are inconsistent with the general trend may be in error. A total of 109 pairs of such erroneous data points are detected and excluded. The final dataset is left with 271 pairs of data points. Multiple linear regression analysis of the 271 pairs of coexisting garnet and clinopyroxene to fit the above model gave: $a = -1629$ (± 636), $b = 3648.55$ (± 391.72), $c = -6.59$ (± 0.35), $d = 1987.98$ (± 79.81), $e = 17.66$ (± 0.68), $f = -1.076$ (± 0.036) ($R^2 = 0.981$). Regression with a $[\text{Mg no. (Gt)}]^2$ term included did not produce a better fit, so that this term was excluded. Consequently the revised garnet-clinopyroxene geothermometer becomes:

$$T(\text{K}) = \{-1629[X_{Ca}^{Gt}]^2 + 3648.55[X_{Ca}^{Gt}] - 6.59[\text{Mg no. (Gt)}] + 1987.98 + 17.66P(\text{kbar})\}/(\ln K_d + 1.076). \qquad (3)$$

In Fig. 4, the new thermometer [Eq. (3)] is compared with previous formulations by Ellis and Green (1979) [Fig. 4a] and Krogh (1988) [Fig. 4b] by plotting the difference between computed temperatures by each method for synthetic garnet-clinopyroxene pairs and the experimental temperatures against the relevant experimental run temperatures. Equation (3) reproduces all experimental run temperatures within ± 100 °C, and gives more accurate temperature estimates, with less scatter for temperatures below 1200 °C. Both the Ellis-Green and Krogh expressions tend to overestimate temperatures below 1200 °C. Whereas estimates above 1200 °C for all three methods are similar, the Krogh expression is most discrepant and the Ellis-Green expression shows a trend of underestimation with increasing experimental temperature (Fig. 4a, b).

Applications

Equation (3) is constructed on the basis of a large amount of experimental data over a wide range of P, T, and composition, therefore it should be applicable to rocks of wide origins and may be successfully extrapolated to natural environments below the experimental temperature range. This is tested in Table 3 by calculation of equilibrium temperatures for rocks from the lower crust (amphibolites, granulites, crustal eclogites) and the upper mantle (eclogite and lherzolite xenoliths in kimberlites, mineral inclusions in diamonds). The new thermometer generally gives more realistic temperature estimates than previously published garnet-clinopyroxene thermometers. For instance, when applied to the metamorphic suites listed by Pattison and Newton (1989), the new thermometer not only produces reasonable temperatures, but also gives narrower ranges of the estimated temperatures for each suite than the Pattison-Newton formulation. Other thermometers (e.g. Krogh 1988; Powell 1985; Ellis and Green 1979) tend to yield temperatures outside the range expected for the corresponding metamorphic grade.

Krogh (1988) calculated equilibrium temperatures in the range of 690–790 °C for a suite of eclogitic rocks from the Tromsø nappe complex within the northern Scandinavian Caledonides. Equation (3) suggests a relatively lower equilibrium temperature range of 636–740 °C, which may be more appropriate for these rocks because eclogites generally have lower equilibrium temperatures than granulites and temperatures higher than 700 °C are normally considered to belong to the granulite facies.

Application of Eq. (3) to four diamondiferous eclogites from southern Africa (Smith et al. 1989), one diamondiferous lherzolite from Sloan, USA (McCallum and Eggler 1976), and two diamondiferous lherzolites from Finsch, South Africa (Shee et al. 1982) generates equilibrium temperatures consistent with diamond stability (Table 3).

Unreliable estimate of Fe^{3+} content in garnet and clinopyroxene, and non-equilibrium between garnet and clinopyroxene are two major sources of errors in estimating equilibrium temperatures of natural rocks. For example, Koons (1984) applied the Ellis-Green thermometer to a suite of eclogites from the Sesia zone of the western Alps, Italy and calculated their temperatures

Table 2. New experimental data on Fe^{2+}/Mg exchange between garnet and clinopyroxene[a]

Run no.	P (kbar)	T °C	X_{Ca}^{Cpx}	Mg no. (cpx)	Mg no. (gt)	X_{Ca}^{Gt}	ln K_d
T3239	20	1200	0.7708	89.68	82.79	0.1546	0.5908
T3101	25	1200	0.7883	89.05	82.71	0.1589	0.5312
T3168	25	1200	0.7242	89.95	83.91	0.1601	0.5409
T3126	27	1200	0.7331	88.86	83.08	0.1507	0.4856
T3454	28	1200	0.8144	90.41	83.32	0.1528	0.6351
T3151	30	1200	0.7700	89.36	82.01	0.1466	0.6106
T3306	35	1200	0.7972	90.38	82.51	0.1377	0.6883
T3355	26	1300	0.7576	89.69	84.16	0.1503	0.4926
T3136	27	1300	0.7250	89.01	83.72	0.1539	0.4543
T3247	27	1300	0.6423	88.88	84.17	0.1468	0.4071
T3364	28	1300	0.7492	89.63	83.20	0.1535	0.5571
T3160	30	1300	0.6432	90.26	86.16	0.1494	0.3981
T3345	35	1300	0.7156	89.56	83.65	0.1416	0.5174
T3167	27	1400	0.6377	88.29	84.28	0.1494	0.3406
T3453	28	1400	0.5393	88.64	85.94	0.1503	0.2444
T3526	29	1400	0.6431	89.03	85.22	0.1430	0.3418
T3163	30	1400	0.6120	88.72	85.05	0.1527	0.3239
T3342	35	1400	0.6224	89.02	83.59	0.1386	0.4648
T3299	35	1400	0.5989	89.49	85.86	0.1450	0.3383
T3378	28	1500	0.4486	89.17	86.05	0.1455	0.2885
T3249	30	1500	0.5251	89.72	87.07	0.1468	0.2594
T3430	30	1500	0.4908	89.20	86.53	0.1530	0.2521
T3527	33	1500	0.4665	89.20	86.25	0.1406	0.2752
T3302	35	1500	0.4961	90.36	86.92	0.1414	0.3441
T3275	27.5	1200	0.7908	89.86	81.68	0.1725	0.6873
T3261	30	1200	0.8081	90.02	81.89	0.1679	0.6909
T3548	32	1200	0.7840	90.01	82.02	0.1744	0.6808
T3269	35	1200	0.8113	90.03	83.38	0.1741	0.6586
T3446	28	1300	0.7357	89.48	82.78	0.1727	0.5702
T3423	30	1300	0.7223	89.26	83.27	0.1789	0.5126
T3556	32	1300	0.7519	90.23	82.90	0.1821	0.6445
T3402	35	1300	0.7567	89.71	82.13	0.1748	0.6408
T3451	32	1400	0.6101	89.27	84.14	0.1661	0.4504
T3417	35	1400	0.6531	89.90	84.26	0.1789	0.5080
T3517	32	1500	0.5211	89.64	85.73	0.1705	0.3652
T3429	34	1500	0.5114	89.32	85.88	0.1622	0.3183
T3268	35	1500	0.4802	89.79	84.62	0.1682	0.4695
T3573	32	1200	0.7982	89.76	81.40	0.1664	0.6953
T3528	35	1200	0.8094	90.01	81.34	0.1795	0.7263
T3583	32	1300	0.7431	89.16	81.96	0.1731	0.5935
T3562	35	1300	0.7260	89.37	81.67	0.1742	0.6357
T3593	38	1300	0.7157	89.73	81.66	0.1861	0.6737
T3568	32	1400	0.6331	88.84	83.20	0.1755	0.4751
T3604	33.5	1400	0.6212	88.82	83.30	0.1722	0.4654
T3536	35	1400	0.6058	88.77	83.40	0.1729	0.4529
T3534	35	1500	0.4643	88.63	84.69	0.1623	0.3428
T3598	38	1500	0.4635	89.31	84.50	0.1740	0.4264

[a] Data from Ai (1992)

ranging from 400 to 1040 °C. He suggested that the spread was caused by the high-jadeite pyroxene solid solutions. Equation (3) gives an even larger temperature range (316–1065 °C) for these rocks. However, if all Fe is allocated to Fe^{2+} in both garnets and clinopyroxenes, the calculated temperatures for half of the rocks reported by Koons (1984) fall in the range of 604–649 °C. Within the other half, two have relatively low calculated temperatures (475 and 384 °C) and two have relatively high calculated temperatures (1083 and 817 °C, Table 3). There is no clear relation between the calculated temperatures and the Na_2O content of the clinopyroxenes. Koons (1984) mentioned that these garnets are usually zoned so the anomalously wide range of temperature estimates may also be a result of non-equilibrium between garnet and clinopyroxene. A suite of eclogite from Cima di Gagnone, Switzerland (Evans et al. 1979) provides another example. The garnets in this suite are zoned, but the clinopyroxenes are not, except in one metarodingite (163-M). If the unzoned clinopyroxene compositions and the garnet rim compositions in samples other than 163-M are used to calculate the equilibrium temperature, a range of 657–716 °C is obtained from Eq. (3) (Table 3). If the garnet core compositions are used, the calculated temperatures range from 582 to 864 °C. In the metarodingite (163-M) the garnet is not zoned but the clinopyroxene is. The clinopyroxene rim composition corresponds to a calculated temperature of 1343 °C; and the core composition,

Table 3. Results of calculated equilibrium temperatures for a range of natural rocks

Sample	ln K_d	X_{Ca}^{Gt}	Mg. no. (gt)	No. of samples	P (kbar)	T3(°C)	TK(°C)	TP(°C)	T(EG) °C	T(PN) °C	T(BM) °C
Amphibolite and granulite (Pattison and Newton 1989)											
Mica Creek	2.13–2.30	0.29–0.37	12.07–18.97	5	6	596–693	645–744	663–765	682–779	521–654	
Eastern Dalradian	1.63–2.26	0.24–0.32	15.38–18.84	6	8	631–756	680–799	697–806	714–821	583–752	
Kapuskasing	1.25–2.13	0.18–0.30	13.64–42.86	11	6.5	554–808	590–894	635–904	657–917	517–758	
Central Gneiss Belt	1.50–1.73	0.17–0.22	16.22–32.05	8	10	707–971	727–986	752–1027	771–1031	625–777	
Adirondack	1.12–2.00	0.19–0.21	13.51–36.99	11	8	595–894	638–951	670–957	691–967	568–916	
Eclogite (Tromsø, North Norway; Krogh 1988)											
	1.50–2.26	0.15–0.44	17–55	17	15	631–740	688–790	705–832	725–844	428–783	
Eclogite (Sesia Zone of the Western Alps, Italy; Koons 1984)											
(With Fe^{3+})	0.87–3.51	0.13–0.30	12.12–28.15	8	15	316–1065	324–987	381–1024	406–1035	220–1330	
(Without Fe^{3+})	0.85–3.04	0.13–0.30	12.06–28.13	8	15	384–1083	387–1002	445–1037	471–1047	287–1362	
Eclogite (Cima di Gagnone, Ticino, Switzerland; Evans et al. 1979)											
(Gt. core)	1.88–2.90	0.25–0.43	25.25–46.68	4	25	582–864	608–899	636–929	655–941		
(Gt. rim)	1.82–2.49	0.17–0.42	37.56–53.61	4	25	657–716	693–766	734–779	757–796		
163-M. Cpx rim	1.34	0.77	56.02	1	25	1343	1066	1501	1481		
163-M. Cpx core	2.02	0.77	56.02	1	25	987	798	1178	1176		
All Fe in garnet allocated to Fe^{2+}											
163-M. Cpx rim	2.62	0.71	40.98	1	25	789	670	937	944		
163-M. Cpx core	1.95	0.71	40.98	1	25	1028	862	1161	1159		
Diamondiferous eclogite (southern Africa; Smith et al. 1989)											
EX2	1.09	0.24	59.08	1	50	1237	1196	1161	1172		
EX10	1.14	0.27	55.45	1	50	1247	1204	1165	1175		
JJG144	1.25	0.19	51.67	1	50	1088	1027	1030	1048		
XM23	1.19	0.23	52.74	1	50	1180	1129	1105	1119		
Diamondiferous lherzolite (Sloan, Wyoming, USA; McCallum and Eggler 1976)											
S1L102	0.96	0.13	83.67	1	51[a]	1091	1063	1097	1113		1056
Diamondiferous lherzolite (Finsch, South Africa; Shee et al. 1982)											
XM46	1.02	0.15	83.94	1	63[a]	1185	1125	1138	1153		1173
XM48	1.06	0.16	82.83	1	57[a]	1126	1103	1110	1126		1089
Diamond inclusions (Sloan, Wyoming, USA; Otter and Gurney 1989)											
A37	1.28	0.26	44.33	1	50	1176	1113	1087	1101		
A73	1.16	0.19	65.02	1	50	1106	1077	1072	1088		
1–10	1.03	0.16	55.84	1	50	1171	1085	1097	1113		
1–15(1)	0.61	0.09	66.23	1	50	1357	1192	1250	1259		
1–15(2)	1.15	0.09	78.34	1	50	929	884	967	990		
Diamond inclusions (Koffiefontein, South Africa; Rickard et al. 1989)											
K2a	0.79	0.10	87.41	1	50	1141	1093	1152	1167		
K8a	1.39	0.22	62.56	1	50	1013	1000	997	1015		
K16a	1.72	0.21	36.41	1	50	922	864	875	897		
K41a	0.67	0.09	49.09	1	50	1361	1142	1209	1221		
K42a	0.74	0.12	45.87	1	50	1370	1177	1208	1219		
K43a	0.65	0.09	63.21	1	50	1325	1153	1220	1231		
K56a	0.97	0.17	68.14	1	50	1199	1152	1147	1160		
Diamond inclusions (Western Australia; Jaques et al. 1989)											
Argyle	0.69–1.30	0.28–0.36	34.86–66.11	6	50	1281–1697	1213–1621	1184–1520	1192–1508		
E4/17	1.17	0.22	52.40	1	50	1174	1116	1097	1112		

T3, equation (3) of this study; TK, Krogh (1988); TP, Powell (1985); T(EG), Ellis and Green (1979); T(PN), Pattison and Newton (1989); T(BM), the two-pyroxene thermometer of Bertrand and Mercier (1985)

[a] Pressures are calculated from the barometer of Nickel and Green (1985). Others are assumed

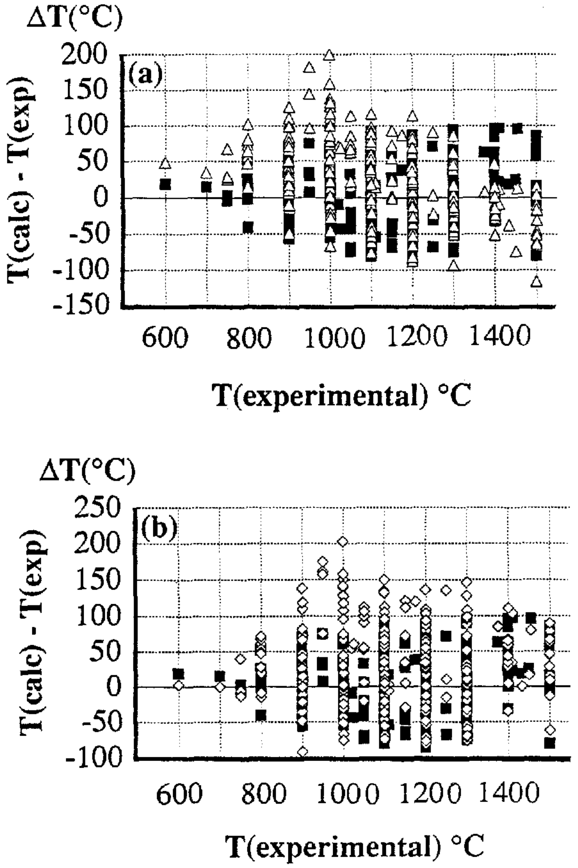

Fig. 4. Comparison of temperatures calculated from Eq. (3) of this study, *solid squares*, the Ellis-Green thermometer, *empty triangles* in (a), and the Krogh formulation, *empty diamonds*, in (b) with the experimental run temperatures. This is shown by the difference between calculated temperatures from each thermometer and the experimental run temperatures. Equation (3) reproduces experimental run temperatures within 100 °C. Both Ellis-Green and Krogh formulations tend to overestimate the experimental run temperatures (in the temperature range of 900 and 1300 °C)

987 °C. Both are significantly higher than the estimates for other samples, which is a direct consequence of the high Fe^{3+} given to the garnet by Evans et al. (1979). Assuming no Fe^{3+} in the garnet, the calculated temperatures for the corresponding clinopyroxene rim and core compositions are 789 and 1028 °C (Table 3). It is noted that the garnet in 163-M contains exceptionally high Fe^{3+} compared with other samples from the same locality and that the given cation total per 12 oxygens for this garnet in 8.053.

Mineral inclusions in single diamonds may not necessarily be in mutual chemical equilibrium. If not, the calculated temperatures may be abnormal. This is best illustrated by Otter and Gurney (1989) and Rickard et al. (1989) and further elaborated by Gurney (1989). Equation (3) is employed to estimate equilibrium temperatures for eclogitic inclusions in diamonds from three localities (Sloan, USA, Otter and Gurney 1989; Koffiefontein, South Africa, Rickard et al. 1989; and Western Australia, Jaques et al. 1989). Most these calculated temperatures are within the range of 1013–1370 °C, with two exceptionally low values (one in the Sloan diamonds, one in the Koffiefontein diamonds) and one extremely high value (in the Western Australia diamonds). These exceptions are probably a result of disequilibrium between garnet and clinopyroxene.

Most calculated temperatures from Eq. (3) are in accordance with estimates from other constraints [e.g. (1) phase relations such as the metamorphic grade and the presence of kyanite; (2) other thermometers based on different reaction schemes such as the two-pyroxene thermometer]. The improvement of the new thermometer over previously published garnet-clinopyroxene thermometers lies in the low temperature range (below 1200 °C) and for rocks with low Mg no. of garnet (mostly high grade metamorphics). For high pressure mantle rocks with temperature above 1200 °C, the new thermometer produces similar temperature estimates to the Ellis and Green thermometer.

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