# Precise Mössbauer milliprobe determination of ferric iron in rock-forming minerals and limitations of electron microprobe analysis

5 authors, including:

Catherine Mccammon
University of Bayreuth
513 PUBLICATIONS   11,512 CITATIONS

SEE PROFILE

Gittel Snyder
New Jersey City University
171 PUBLICATIONS   3,494 CITATIONS

SEE PROFILE

Some of the authors of this publication are also working on these related projects:

Materials at high-P,T View project

Silicates at high pressures View project

# Precise Mössbauer milliprobe determination of ferric iron in rock-forming minerals and limitations of electron microprobe analysis

VLADIMIR N. SOBOLEV,[1,3,*] CATHERINE A. MCCAMMON,[2] LAWRENCE A. TAYLOR,[1]
GREGORY A. SNYDER,[1] AND NIKOLAI V. SOBOLEV[3]

[1]Planetary Geosciences Institute, Department of Geological Sciences, University of Tennessee,
Knoxville, Tennessee 37996, U.S.A.
[2]Bayerisches Geoinstitut, Universität Bayreuth, D-95440 Bayreuth, Germany
[3]Institute of Mineralogy and Petrography, Siberian Branch of the Russian Academy of Sciences, 630090 Novosibirsk, Russia

## ABSTRACT

For estimations of *P-T* conditions of igneous and metamorphic rocks, $Fe^{3+}$ in coexisting minerals is either assumed to be zero or is calculated from electron microprobe analyses (EMPA) based upon stoichiometry and charge balance. Geothermobarometers that involve $Fe^{2+}$-$Mg^{2+}$ exchange can be significantly affected by either neglecting $Fe^{3+}$ or using incorrect values. Ratios of $Fe^{3+}/\Sigma Fe$ in garnet and clinopyroxene measured by a Mössbauer milliprobe were compared to those calculated from EMPA of garnet and clinopyroxene from eclogite xenoliths from the Udachnaya kimberlite in Yakutia. The effects of $Fe^{3+}$ contents in garnet and clinopyroxene on temperature estimations were evaluated.

The following $Fe^{3+}/\Sigma Fe$ (in at%) values were obtained (EMPA/Mössbauer): Gt = 9.4/6.0; 11.5/7.0; 19.4/16.0; and 24.7/15.0; Cpx = 22.0/22.9; 34.2/22.0. The effects of $Fe^{3+}$ in clinopyroxenes on calculated temperatures are illustrated by taking eclogitic clinopyroxene compositions and changing contents of certain elements within the range of standard deviations for EMPA of those particular elements. Increasing $Na_2O$ contents from 5.67 to 5.74 wt% (<2.0% relative error) would lead to increasing $Fe^{3+}/\Sigma Fe$ from 31.6 to 47.1%, thereby decreasing the calculated temperature from 1026 to 941 °C. Various $Fe^{3+}/\Sigma Fe$ values for garnet and clinopyroxene were also tested for their effects on calculated temperatures: for clinopyroxene, *T* decreases with increasing $Fe^{3+}/\Sigma Fe$ whereas for garnet, *T* increases with increasing $Fe^{3+}/\Sigma Fe$. This compensation effect between garnet and clinopyroxene moderates the variation in temperature estimations of eclogites based on $Fe^{3+}$ corrected vs. uncorrected microprobe analyses.

Little correlation exists between EMPA-calculated and Mössbauer-measured $Fe^{3+}/\Sigma Fe$ values for these mantle-derived garnets and clinopyroxenes. Even a small relative error in $Fe^{3+}$ may significantly change calculated temperatures of equilibration, seriously affecting petrologic interpretations. In particular, uncertainty in $Fe^{3+}$ calculated from EMPA of silicate minerals leads to serious questions with regard to $K_D$ values obtained from natural assemblages.

## INTRODUCTION

Several chemical elements can occur in multiple oxidation states in both terrestrial and extraterrestrial minerals. Iron is one example of a multivalent element that is a common constituent of silicate minerals in igneous and metamorphic rocks. The significance of variation of $Fe^{3+}/(Fe^{3+} + Fe^{2+})$ is difficult to overestimate. Many petrogenetic and thermodynamic parameters of Fe-bearing minerals involve $Fe^{2+}$, such as $X_{mg} = (Mg/[Mg + Fe^{2+}])$ and $K_D$, the $Fe^{2+}$-Mg distribution coefficient. These values are a direct function of $Fe^{2+}$ and $Fe^{3+}$ contents. However, in most cases, modern analytical techniques are incapable of measuring oxidation state, resulting in various as-sumptions (e.g., that all Fe present in a sample has the same valence).

Previously, there were only two analytical methods commonly used to determine $Fe^{3+}/\Sigma Fe$ and $Fe_2O_3$ content directly, Mössbauer spectroscopy, and wet chemistry. These methods permit distinction between $Fe^{2+}$ and $Fe^{3+}$, which is crucial to any estimation of redox and *P-T* conditions during mineral formation. However, both techniques require relatively large amounts of sample ($\geq 50$ mg for Mössbauer spectroscopy for a sample containing 10 wt% FeO, and $\geq 0.5$ g for wet chemistry). These mass requirements have made it impossible to analyze single mineral grains or minerals in thin section (i.e., in situ). Recently, another technique for $Fe^{3+}/\Sigma Fe$ measurement has become available, X-ray absorption near-edge struc-

ture (XANES) spectroscopy (Delaney et al. 1996), which permits the measurement of $Fe^{3+}/\Sigma Fe$ in-situ in $50 \times 30$ $\mu m$ areas.

It has been common practice to use the crystal chemistry of silicate phases analyzed by the electron microprobe (EMP) to calculate the $Fe^{3+}$ content of a mineral. However, this can result in gross imprecision of $Fe^{3+}/\Sigma Fe$ determinations and, thus, large uncertainties in subsequent $P$-$T$ estimations. Attempts have been made to compare and correlate measured and calculated $Fe^{3+}$ contents in silicates from upper mantle xenoliths. Dyar et al. (1984) showed that $Fe^{3+}$ contents measured using conventional Mössbauer spectroscopy are different from those calculated from EMP analyses of spinels, clino-, and orthopyroxenes. Wood and Virgo (1989) used Mössbauer-analyzed spinels as secondary EMP standards and stated that this technique improves the precision dramatically, approaching $\pm 0.002$ $X_{Fe3O4}$ in analyzed spinels. These authors concluded that for routine EMP analysis of spinel, the largest systematic uncertainties are associated with the concentration of $Al_2O_3$. Luth et al. (1990) stated that for a peridotite mineral assemblage (i.e., garnet, olivine, clino-, and orthopyroxene), the values of $Fe^{3+}/\Sigma Fe$ measured by the Mössbauer technique and those calculated from EMP analyses were significantly different. They also determined a positive correlation between $Fe^{3+}/\Sigma Fe$ contents in garnet and temperature estimates.

Several studies have attempted to assess the precision of $Fe^{2+}/Fe^{3+}$ determination by EMP analyses. The uncertainty in determination of formula proportions of cations (and $Fe^{3+}$ in particular) from EMP analyses was addressed by Finger (1972) who showed that $Fe^{3+}$ in chromite can be determined with a precision of $\pm 12\%$ (0.534 $\pm$ 0.065 $Fe^{3+}$). Canil and O'Neill (1996) used the method of Dollase and Newman (1984) to show that the level of imprecision in $Fe^{3+}$ determined from EMP analyses is related to the total Fe content, and increases in the order spinel<garnet<clinopyroxene. The level of imprecision for clinopyroxenes may be too large to be useful petrologically (Canil and O'Neill 1996). Moreover, even minimal errors in determination of high-valence and high-abundance elements such as Si in clinopyroxene can significantly affect formula proportions of cations calculated from EMP analyses.

Recent advances in the Mössbauer technique, in particular the development of the Mössbauer milliprobe, have made it possible to perform in situ measurements of $Fe^{3+}/\Sigma Fe$ in single mineral grains on a routine basis (McCammon 1994). For example, McCammon et al. (1997) determined $Fe^{3+}$ in garnets and clinopyroxenes inclusions ($\sim$200 $\mu m$ in size) within diamond from the George Creek kimberlite. Their results showed that there is little correlation between $Fe^{3+}/\Sigma Fe$ determined by the EMP analyses and Mössbauer techniques for these inclusions.

We have used the Mössbauer milliprobe to determine $Fe^{3+}/\Sigma Fe$ in silicates from eclogites from the Udachnaya kimberlite, Yakutia. Eclogites are basically bimineralic

rocks consisting of the primary minerals pyropic garnet and omphacitic clinopyroxene. Therefore, only geothermometers utilizing the garnet-clinopyroxene Fe-Mg exchange are applicable to this type of rock (e.g., Råheim and Green 1974; Ellis and Green 1979; Powell 1985; Krogh 1988). Due to the fact that only $Fe^{2+}$ takes part in this cation exchange between garnet and clinopyroxene, determining accurate $Fe^{3+}$ values is extremely important. In this paper, we compare the $Fe^{3+}/\Sigma Fe$ ratios of these minerals measured by the Mössbauer milliprobe to those ratios calculated from EMP analyses. In addition, we evaluate errors in other elements associated with routine EMP analyses with regard to calculated $Fe^{3+}$ values. Last, we address the effects of $Fe^{3+}$ contents in garnet and clinopyroxene on estimates of equilibration temperatures for minerals in eclogite xenoliths.

## ANALYTICAL TECHNIQUES

### Mössbauer analyses

$Fe^{3+}$ was determined using a Mössbauer milliprobe at the Bayerisches Geoinstitut, Universität Bayreuth, Germany. The Mössbauer milliprobe is a conventional Mössbauer spectrometer linked to some additional equipment (full details of the method are given in McCammon et al. 1991). To obtain adequate count rates, the conventional Mössbauer source (typical specific activity 100 mCi/cm²) was replaced by a point source (specific activity $\geq$2000 mCi/cm²). The gamma rays were collimated to the selected sample diameter using a lead shield, and the source-to-sample distance was reduced to <5 mm. This reduction results in a solid angle similar to conventional experiments and, hence, a similar count rate. Because the signal quality depends on absorber density (measured in mg Fe/cm²) and not on the total amount of Fe in the sample, the reduction in sample size has no effect on the effective thickness of the absorber.

The sample mount consisted of an epoxy disk embedded with sample grains approximately 850 $\mu m$ thick and 750 $\mu m$ in diameter. A 500 $\mu m$ diameter hole was drilled in a piece of 25 $\mu m$ thick Ta foil (absorbs 99% of 14.4 keV gamma rays) and placed over the grain to be studied. Samples were run at room temperature, and each set of spectra took 1–4 days to collect.

### Electron microprobe analyses

Major-element compositions were determined by EMPA using the four-spectrometer CAMECA SX-50 in the Department of Geological Sciences at the University of Tennessee. The accelerating voltage was 15 kV, with a beam current of 20 nA for clinopyroxenes and 30 nA for garnets. The beam size was 10 $\mu m$, and counting times of 20 s were used for all elements and backgrounds, except for Na in garnet and K in clinopyroxene (60 s). All data were corrected using ZAF procedures adapted for the CAMECA probe.
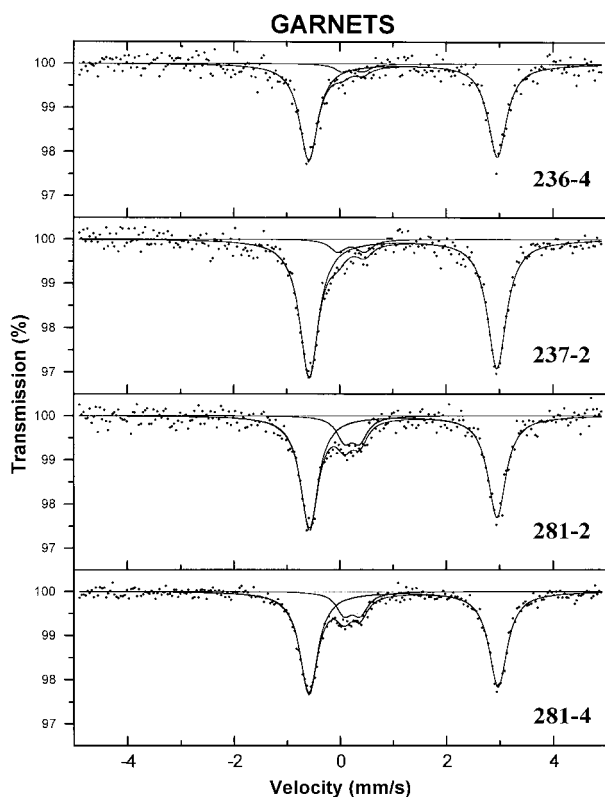
## GARNETS



FIGURE 1. Mössbauer spectra of garnets from samples U-236, U-237, and U-281. Small doublet (shaded) in all spectra is $Fe^{3+}$.

## CLINOPYROXENES



FIGURE 2. Mössbauer spectra of clinopyroxenes from samples U-236 and U-281. Small doublet (shaded) in both spectra is $Fe^{3+}$.

## SAMPLE CHARACTERIZATION

Eclogite xenoliths, from which garnet and clinopyroxene samples were extracted, represent part of an extensive suite of diamondiferous eclogites from the Udachnaya kimberlite pipe, Yakutia, Russia. They were described in detail in previous studies by our group (Jerde et al. 1993; Sobolev et al. 1994; Snyder et al. 1997).

The samples (U-236, U-237, and U-281) are coarse-grained and equigranular, and consist of a bimineralic assemblage of garnet and clinopyroxene. Modal proportions of garnet and clinopyroxene are 64:36 in U-236, 51:49 in U-237, and 60:40 in U-281 (Sobolev et el. 1994). Both garnet and clinopyroxene from these samples are chemically homogeneous, with the exception of U-281 where intra-grain $Cr_2O_3$ variation is present in garnet (0.50 to 1.50 wt% $Cr_2O_3$). All studied samples were classified as Group A eclogites (Sobolev et al. 1994) according to the criteria of Coleman et al. (1965), as modified by Taylor and Neal (1989). Garnets in all three xenoliths are relatively fresh, with <20–30% alteration, whereas clinopyroxenes are 60–70% altered (Sobolev et al. 1994). Fresh garnet and clinopyroxene pairs from each sample (only garnet for U-237) were selected for analyses. All selected grains lacked visible alteration products.
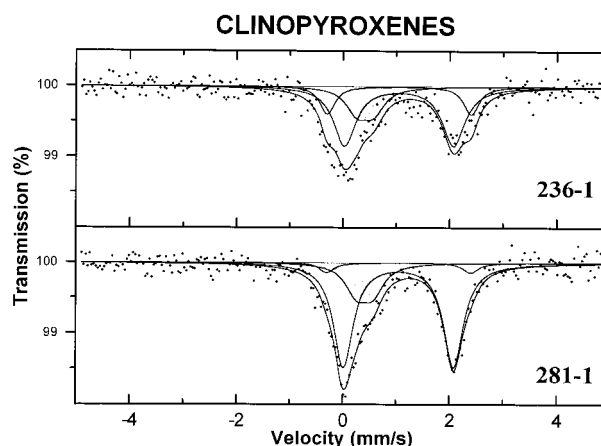
## RESULTS

Mössbauer spectra of Udachnaya garnets and clinopyroxenes are shown in Figures 1 and 2. The garnet spectra were fit to one $Fe^{2+}$ doublet and one $Fe^{3+}$ doublet according to conventional models (e.g., Amthauer et al. 1976). The subspectra are well resolved, and $Fe^{3+}/\Sigma Fe$ can be determined with a high degree of confidence. Inasmuch as garnet is cubic, there should be no area asymmetry of component doublets due to texture effects. Nevertheless, the $Fe^{2+}$ doublet appears to have a slight asymmetry, which is consistent with observations of Geiger et al. (1992); consequently, the area ratio of the $Fe^{2+}$ doublet was allowed to vary. Otherwise, conventional constraints were applied (line widths and areas of components of each doublet were constrained to be 1:1).

The clinopyroxene spectra were fit to two $Fe^{2+}$ doublets (corresponding to M1 and M2 sites) and one $Fe^{3+}$ doublet according to conventional models (e.g., Luth and Canil 1993). Resolution of clinopyroxene spectra is significantly reduced compared to garnet spectra because of the similarity of hyperfine parameters for subspectra. Therefore, the following constraints were applied in addition to the conventional ones of equal line widths and areas of components: (1) the center shift, quadrupole splitting, and line width of the M1 doublet were constrained to average values observed by Luth and Canil (1993). Since the relative abundance of this doublet is small (<20%), errors in the hyperfine parameters of the doublet have little effect on the measured $Fe^{3+}/\Sigma Fe$ ratio; (2) for the U-236-1 clinopyroxene sample, where the signal:noise ratio was not as high, the center shift, quadrupole splitting, and line width of the $Fe^{3+}$ doublet were constrained to the values determined for the U-281-1 clinopyroxene spectrum. This constraint contributes a higher error to the $Fe^{3+}/\Sigma Fe$ determination of this sample.

## COMPARISON OF EMP AND MÖSSBAUER RESULTS

A comparison of EMP and Mössbauer analyses of four garnets and two clinopyroxenes are given in Table 1 and

**TABLE 1.** Garnet and clinopyroxene major-element compositions determined by electron microprobe

| | Garnet | | | | | | | | Clinopyroxene | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | 236-4* E | 236-4 M | 237-2 E | 237-2 M | 281-2 E | 281-2 M | 281-4 E | 281-4 M | 236-1 E | 236-1 M | 281-1 E | 281-1 M |
| $SiO_2$ | 40.8 | 40.8 | 40.8 | 40.8 | 40.8 | 40.8 | 40.5 | 40.5 | 54.5 | 54.5 | 54.7 | 54.7 |
| $TiO_2$ | 0.31 | 0.31 | 0.33 | 0.33 | 0.47 | 0.47 | 0.55 | 0.55 | 0.45 | 0.45 | 0.20 | 0.20 |
| $Al_2O_3$ | 22.8 | 22.8 | 22.9 | 22.9 | 21.7 | 21.7 | 20.8 | 20.8 | 5.72 | 5.72 | 1.81 | 1.81 |
| $Cr_2O_3$ | 0.14 | 0.14 | 0.10 | 0.10 | 0.50 | 0.50 | 1.49 | 1.49 | 0.10 | 0.10 | 0.31 | 0.31 |
| $Fe_2O_3$ | 1.17 | 0.75 | 1.50 | 0.91 | 2.61 | 2.16 | 3.34 | 2.03 | 0.62 | 0.63 | 1.58 | 1.02 |
| FeO | 10.2 | 10.6 | 10.4 | 10.9 | 9.78 | 10.2 | 9.18 | 10.4 | 1.92 | 1.91 | 2.74 | 3.25 |
| MnO | 0.24 | 0.24 | 0.30 | 0.30 | 0.48 | 0.48 | 0.49 | 0.49 | 0.02 | 0.02 | 0.11 | 0.11 |
| MgO | 14.8 | 14.8 | 15.9 | 15.9 | 17.6 | 17.6 | 17.4 | 17.4 | 13.6 | 13.6 | 16.3 | 16.3 |
| CaO | 9.35 | 9.35 | 7.56 | 7.56 | 5.77 | 5.77 | 6.20 | 6.20 | 19.7 | 19.7 | 20.6 | 20.6 |
| $Na_2O$ | 0.08 | 0.08 | 0.08 | 0.08 | 0.06 | 0.06 | 0.07 | 0.07 | 3.03 | 3.03 | 1.57 | 1.57 |
| $K_2O$ | ND† | ND | ND | ND | ND | ND | ND | ND | <0.02 | <0.02 | <0.02 | <0.02 |
| $FeO_T$‡ | 11.3 | 11.3 | 11.7 | 11.7 | 12.1 | 12.1 | 12.2 | 12.2 | 2.48 | 2.48 | 4.17 | 4.17 |
| Total | 99.70 | 99.71 | 99.69 | 99.69 | 99.54 | 99.54 | 99.69 | 99.69 | 99.59 | 99.59 | 99.74 | 99.74 |
| Si | 2.974 | 2.976 | 2.964 | 2.967 | 2.964 | 2.967 | 2.948 | 2.957 | 1.963 | 1.963 | 1.986 | 1.988 |
| Ti | 0.017 | 0.017 | 0.018 | 0.018 | 0.026 | 0.026 | 0.030 | 0.030 | 0.012 | 0.012 | 0.005 | 0.005 |
| $^{IV}Al$ | | | | | | | | | 0.037 | 0.037 | 0.014 | 0.012 |
| $^{VI}Al$ | 1.957 | 1.959 | 1.960 | 1.963 | 1.858 | 1.860 | 1.784 | 1.789 | 0.206 | 0.206 | 0.063 | 0.066 |
| Cr | 0.008 | 0.008 | 0.006 | 0.006 | 0.029 | 0.029 | 0.086 | 0.086 | 0.003 | 0.003 | 0.009 | 0.009 |
| $Fe^{3+}$ | 0.064 | 0.041 | 0.082 | 0.050 | 0.143 | 0.118 | 0.183 | 0.112 | 0.017 | 0.017 | 0.043 | 0.028 |
| $Fe^{2+}$ | 0.621 | 0.647 | 0.629 | 0.663 | 0.594 | 0.619 | 0.559 | 0.632 | 0.058 | 0.058 | 0.083 | 0.099 |
| Mn | 0.015 | 0.015 | 0.018 | 0.019 | 0.030 | 0.030 | 0.030 | 0.030 | 0.001 | 0.001 | 0.003 | 0.004 |
| Mg | 1.603 | 1.604 | 1.723 | 1.725 | 1.901 | 1.903 | 1.888 | 1.891 | 0.732 | 0.732 | 0.881 | 0.882 |
| Ca | 0.730 | 0.731 | 0.588 | 0.589 | 0.449 | 0.449 | 0.483 | 0.485 | 0.760 | 0.760 | 0.800 | 0.801 |
| Na | 0.011 | 0.011 | 0.012 | 0.012 | 0.008 | 0.008 | 0.009 | 0.009 | 0.212 | 0.212 | 0.111 | 0.111 |
| K | | | | | | | | | 0.001 | 0.001 | 0.001 | 0.001 |
| Total | 8.000 | 8.008 | 8.000 | 8.011 | 8.000 | 8.008 | 8.000 | 8.024 | 4.000 | 4.000 | 4.000 | 4.005 |
| $X_{Mg}$ | 0.72 | 0.71 | 0.73 | 0.72 | 0.76 | 0.75 | 0.77 | 0.75 | 0.93 | 0.93 | 0.91 | 0.90 |
| $Fe^{3+}/\Sigma Fe$ | 9.4 | 6.0 | 11.5 | 7.0 | 19.4 | 16.0 | 24.7 | 15.0 | 22.6 | 22.9 | 34.2 | 22.0 |
| $T_E$§ (°C) | 953 | | | | 1030 | | | | | | | |
| $T_M$ (°C) | 930 | | | | 1068 | | | | | | | |

*Note:* Fe³⁺/ΣFe ratio is calculated from EMPA ("E" columns) and measured by Mössbauer spectroscopy ("M" columns).
\* Each EMP analysis (E) is the average of 20 points on four grains.
† ND = not determined.
‡ All iron determined as FeO by EMPA.
§ Temperature estimates (Ellis and Green 1979) for samples U-236 and U-281 (4.5 GPa pressure is used).
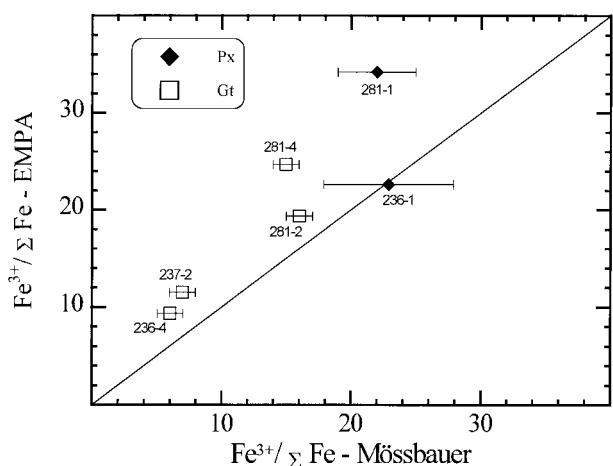


**FIGURE 3.** Comparison of analyzed (Mössbauer) and calculated (EMP) Fe³⁺/ΣFe ratios of garnets and clinopyroxenes from Udachnaya eclogites. Error bars are shown for both garnet and clinopyroxene analyses.

shown in Figure 3. It is evident that, although measured Fe³⁺/ΣFe ratios approach the calculated ones (e.g., Gt 281-2, Cpx 236-1), there is no direct (1:1) correlation between the two sets of data. Note that calculated Fe³⁺ values are always greater than measured ones, except clinopyroxene 236-1, where they are roughly equal. Two different garnets from one sample (U-281) produced no measurable differences in Fe³⁺/ΣFe (Table 1, Fig. 3).

Some Fe³⁺ should be present in almost all igneous and metamorphic Fe-bearing silicate minerals, depending on the $f_{O_2}$ of the system during their formation and equilibration. However, due to the differences in crystal structure and the amounts of total FeO, the Fe³⁺/ΣFe ratio is different in coexisting minerals. The difference in Fe³⁺/ΣFe between garnet and clinopyroxene in two different samples is illustrated by the results in Table 1, where two different Fe³⁺/ΣFe ratios of garnets (6.0 and 15.0) correspond to similar (22.9 and 22.0) Fe³⁺/ΣFe in clinopyroxenes (Table 1, samples 236 and 281).

Values of Fe³⁺ obtained from Mössbauer analyses of clinopyroxenes possess much larger errors than those of garnets (Fig. 3). This error is a result of relatively low FeO contents in clinopyroxene (2.48 wt% in U-236 and 4.17 wt% in U-281). Since the quality of a Mössbauer analysis depends on sample density (measured in mg Fe/

cm²; McCammon et al. 1995), samples with the same thickness but different concentrations of FeO would have different precisions in the determination of $Fe^{3+}/\Sigma Fe$.

Although similar for some minerals (garnet 281-2, clinopyroxene 236-1, Table 1), $Fe^{3+}/\Sigma Fe$ obtained from the EMP and Mössbauer techniques are significantly different in most cases. However, temperatures calculated for eclogites using EMP-determined $Fe^{3+}$ values are not greatly different from those calculated using Mössbauer values, and all are within ±40 °C range (Table 1). The reason for this is discussed below.

## Influence of Fe³⁺ on Fe-Mg distribution in eclogitic minerals and calculated temperatures of equilibration

When calculating *P-T* parameters using Fe-bearing minerals, $Fe^{3+}$ is either assumed to be zero, or it is calculated from EMP analyses based upon stoichiometry and charge balance. As discussed below, the EMP technique does not always provide an accurate $Fe^{3+}$ value, particularly for silicates, which can seriously affect the petrologic interpretation of the results. The importance of separating different oxidation states of iron can be illustrated by looking at certain key petrologic parameters, such as $X_{Mg}$ and the $Fe^{2+}$-Mg distribution coefficient ($K_D$). The distribution coefficient ($K_D$) forms the basis of several geothermometers (e.g., Råheim and Green 1974; Ellis and Green 1979; Powell 1985; Krogh 1988), therefore, the calculated temperature of equilibration strongly depends on it. Unfortunately, there are no experimental data on the effects of $Fe^{3+}$ for existing geothermometers and geobarometers. Luth et al. (1990) showed that using a combination of the O'Neill and Wood (1979) geothermometer and the MacGregor (1974) geobarometer, the two values of $Fe^{3+}/\Sigma Fe$ of 0 and 0.12 would change calculated pressure from 59 to 74 kbar and calculated temperature from 1269 to 1505 °C, respectively. Note that neither of these geothermometers or geobarometers involves clinopyroxene, which may contain significant amounts of $Fe^{3+}$. It is obvious that those geothermometers and geobarometers that involve $Fe^{2+}$-Mg exchange would be most significantly affected by either neglecting $Fe^{3+}$ or using incorrect values.

### Electron microprobe calculated Fe³⁺

The effect of $Fe^{3+}$ in clinopyroxene on calculated temperatures can be illustrated by taking typical eclogitic clinopyroxene compositions (sample U-107/2, data from Sobolev et al. 1994) and clinopyroxene compositions from a Gt-Cpx intergrowth (sample A-143, data from Sobolev et al. 1997) and changing slightly the abundances of Si and Na within the range given by the analytical error, i.e., standard deviation (Table 2). Note that the error in $Fe^{3+}/\Sigma Fe$ is controlled largely by Si contents in garnet and clinopyroxene, due to its major abundance and +4 charge. In addition, eclogitic clinopyroxenes contain relatively high Na contents (up to 10.7 wt%, Sobolev 1974). Because of its low atomic number, sodium requires ad-

ditional precision when analyzing clinopyroxene. The +1 charge of Na makes the error in Na analysis very sensitive when calculating formula proportions of cations.

We used "auoestimated errors in EMP analyses"calculated by Canil and O'Neill (1996), which is based on the method proposed by Dollase and Newman (1984). As explained by Dollase and Newman (1984), the statistical validity of the method depends on a knowledge of the standard deviations of the measured concentrations of each oxide component. These errors were calculated with an assumption that the EMP analyses contain no systematic errors, so that these standard deviations result from counting statistics alone. With this in mind, we changed the $SiO_2$ and $Na_2O$ contents of the clinopyroxenes as shown in Table 2. Although the errors proposed by Canil and O'Neill (1996) are optimistic, even these very conservative relative errors in Na (1.2%) and in Si (0.7%) analyses would drastically change the $Fe^{3+}/\Sigma Fe$ ratio and the estimated temperature of equilibration (Table 2). These changed values for $SiO_2$ and $Na_2O$ are well within the range of precision for clinopyroxene compositions and reflect a real EMP analysis. Note that the temperature for sample A-143, estimated by assuming that $Fe^{3+}/\Sigma Fe = 0$, is 1544 °C (Table 2). This is 540 °C higher than the temperature estimated by using the $Fe^{3+}$ calculated from the EMP analysis.

The lack of precision of EMP analyses also can lead to large errors in calculating formula proportions of cations. A method of estimating uncertainties associated with the coefficients of the mineral formula was proposed by Finger (1972). This was shown by assuming an error in EMP analyses of chromite of 2%, and the uncertainties for major elements were within 0.5–2% range. At the same time, the uncertainty in the $Fe^{3+}$ contents was 12%. This method, however, may not be useful with eclogitic minerals where $Fe^{3+}/\Sigma Fe$ ratios seldom exceed values of 30.0 [compared to 62.0 in chromite (Finger 1972)]. The reproducibility (precision) in EMP analyses combined with inter- and intragranular compositional variations can also result in large errors in temperature estimates, on the order of ±100 °C or more. Thus, it is difficult to estimate an accurate temperature of equilibration since it may lie within a 200 °C range. The error may be even larger if an energy dispersive spectrometer (EDS) is used for quantitative analyses (e.g., Jacob et al. 1994), particularly since even small changes in $Na_2O$ can have a significant affect on calculated $Fe^{3+}$.

### Effect of $K_D$ on Gt-Cpx thermometers

Although included as part of the $Fe^{2+}$-Mg exchange geothermometers, $K_D$ itself is not significantly affected by the imprecision in $Fe^{3+}$ estimation when values from EMP and Mössbauer analyses are compared. We used the data from Tables 1 and 2 and plotted Gt-Cpx pairs on a $f_{Gt}$-$f_{Cpx}$ graph, where f is $Fe^{2+}/(Fe^{2+} + Mg)$ of the mineral (in mol%). There is almost no difference in f (Gt and Cpx) and $K_D$ values in the sample 236 between EMP-calculated and Mössbauer-measured $Fe^{2+}/Fe^{3+}$ ratios (Fig.

**TABLE 2.** Garnet and clinopyroxene major-element compositions of diamondiferous eclogite U-107/2 (Udachnaya kimberlite, Soboley et al. 1994) and garnet peridotite A-143 (Mir kimberlite, Soboley et al. 1997)

| | U-107/2 | | | | | A-143 | | | |
|---|---|---|---|---|---|---|---|---|---|
| | Gt | Cpx | Cpx-1 | Cpx-2 | Cpx-3 | Gt | Cpx | Cpx-4 | Cpx-5 |
| $SiO_2$ | 40.3 | 55.1 | 54.8 | 55.1 | 54.8 | 41.8 | 55.1 | 55.1 | 55.1 |
| $TiO_2$ | 0.33 | 0.16 | 0.16 | 0.16 | 0.16 | 0.07 | 0.14 | 0.14 | 0.14 |
| $Al_2O_3$ | 22.3 | 10.3 | 10.3 | 10.3 | 10.3 | 19.8 | 5.92 | 5.92 | 5.92 |
| $Cr_2O_3$ | 0.05 | 0.04 | 0.04 | 0.04 | 0.04 | 4.98 | 6.09 | 6.09 | 6.09 |
| $Fe_2O_3$ | 0.94 | 0.81 | 1.21 | 1.28 | 1.57 | 1.19 | 2.46 | 2.87 | 1.99 |
| FeO | 9.75 | 1.58 | 1.22 | 1.16 | 0.90 | 7.31 | 1.26 | 0.89 | 1.68 |
| MnO | 0.19 | 0.03 | 0.03 | 0.03 | 0.03 | 0.53 | 0.13 | 0.13 | 0.13 |
| MgO | 11.2 | 10.4 | 10.4 | 10.4 | 10.4 | 21.7 | 10.7 | 10.7 | 10.7 |
| CaO | 14.1 | 15.1 | 15.1 | 15.1 | 15.1 | 2.49 | 11.4 | 11.4 | 11.4 |
| $Na_2O$ | 0.11 | 5.67 | 5.67 | 5.74 | 5.74 | 0.08 | 6.71 | 6.79 | 6.63 |
| $K_2O$ | ND | 0.08 | 0.08 | 0.08 | 0.08 | ND | <0.02 | <0.02 | <0.02 |
| $FeO_{tot}$ | 10.6 | 2.31 | 2.31 | 2.31 | 2.31 | 8.38 | 3.47 | 3.47 | 3.47 |
| Total | 99.18 | 99.19 | 98.89 | 99.28 | 98.96 | 99.83 | 99.59 | 99.67 | 99.50 |
| Si | 2.988 | 1.966 | 1.960 | 1.963 | 1.958 | 2.995 | 1.985 | 1.982 | 1.988 |
| Ti | 0.018 | 0.004 | 0.004 | 0.004 | 0.004 | 0.004 | 0.004 | 0.004 | 0.004 |
| $^{IV}Al$ | | 0.034 | 0.040 | 0.037 | 0.042 | | 0.015 | 0.018 | 0.012 |
| $^{VI}Al$ | 1.948 | 0.399 | 0.394 | 0.396 | 0.391 | 1.667 | 0.236 | 0.233 | 0.240 |
| Cr | 0.003 | 0.001 | 0.001 | 0.001 | 0.001 | 0.282 | 0.173 | 0.173 | 0.174 |
| $Fe^{3+}$ | 0.053 | 0.022 | 0.033 | 0.031 | 0.042 | 0.064 | 0.067 | 0.078 | 0.054 |
| $Fe^{2+}$ | 0.605 | 0.047 | 0.037 | 0.037 | 0.027 | 0.438 | 0.038 | 0.027 | 0.051 |
| Mn | 0.012 | 0.001 | 0.001 | 0.001 | 0.001 | 0.032 | 0.004 | 0.004 | 0.004 |
| Mg | 1.238 | 0.553 | 0.555 | 0.552 | 0.554 | 2.316 | 0.572 | 0.571 | 0.573 |
| Ca | 1.120 | 0.577 | 0.579 | 0.576 | 0.578 | 0.191 | 0.438 | 0.437 | 0.439 |
| Na | 0.016 | 0.392 | 0.393 | 0.397 | 0.398 | 0.011 | 0.469 | 0.473 | 0.463 |
| K | | 0.004 | 0.004 | 0.004 | 0.004 | | 0.000 | 0.000 | 0.000 |
| Total | 8.000 | 4.000 | 4.000 | 4.000 | 4.000 | 8.000 | 4.000 | 4.000 | 4.000 |
| Total* | 8.018 | 4.007 | 4.011 | 4.011 | 4.011 | 8.021 | 4.022 | 4.026 | 4.018 |
| $X_{Mg}$ | | 0.92 | 0.94 | 0.94 | 0.94 | | 0.94 | 0.96 | 0.92 |
| $Fe^{3+}/\Sigma Fe$† | 8.0 | 31.6 | 47.1 | 45.7 | 61.1 | 12.7 | 63.7 | 74.4 | 51.6 |
| T °C | | 1026 | 941 | 949 | 851 | 1004 | | 870 | 1143 |

*Note:* Temperatures were estimated with the geothermometer of Ellis and Green (1979); 4.5 GPa pressure was used. An error of 0.7% of the amount present (i.e., 0.3 wt% $SiO_2$) was subtracted from the real analysis (Cpx-1); an error of 1.2% of the amount present (i.e., 0.07 wt% $Na_2O$) was added (Cpx-2) for the sample U-107/2; both of the errors were applied (Cpx-3). An error of 0.08 wt% $Na_2O$ is added (Cpx-4) and subtracted (Cpx-5) for the sample A-143. Relative precisions of 0.75% for $SiO_2$ and 1.5% for $Na_2O$ are optimistic but are values used by Canil and O'Neil (1996).
  * Before calculating $Fe^{3+}$.
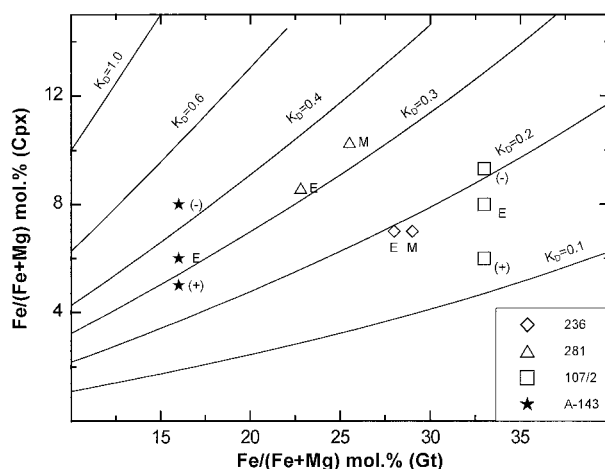  † Calculated from EMPA.



**FIGURE 4.** $Fe^{2+}/(Fe^{2+} + Mg)_{Gt}$ ($f_{Gt}$) vs. $Fe^{2+}/(Fe^{2+} + Mg)_{Cpx}$ ($f_{Cpx}$) for Udachnaya eclogites 236, 281, 107/2, and Mir peridotite A-143 (f = $Fe^{2+}/[Fe^{2+} + Mg]$ in mol%). E = EMP calculated; M = Mössbauer analyzed; (−) for star = subtracted 0.08 wt% of $Na_2O$; (−) for square = subtracted 0.07 wt% of $Na_2O$; (+) for star = added 0.08 wt% of $Na_2O$; (+) for square = added 0.07 wt% of $Na_2O$ (see detailed explanation in Table 2 and text).

4). Although f values for garnet and clinopyroxene in sample 281 are different, the distribution coefficients are the same (Fig. 4).

The situation changes when clinopyroxene compositions with high $Na_2O$ (>4–5 wt%) are considered. The distribution coefficient varies from 0.14 (+0.07 wt% of $Na_2O$) to 0.21 (−0.07 wt% of $Na_2O$) in sample 107/2 (Fig. 4), and from 0.25 (+0.08 wt% of $Na_2O$) to 0.47 (−0.08 wt% of $Na_2O$) in sample A-143 (Fig. 4). Therefore, a potential error in EMP analyses of high-Na clinopyroxenes may lead to a wide variation of Fe-Mg distribution coefficient as well as to significant differences in calculated temperatures.

It is clear that $K_D$ is not only a function of temperature. It has been shown that pressure as well as garnet composition may affect Fe-Mg distribution coefficient in eclogites (e.g., Sobolev 1964; Banno 1970; Ai 1994). In fact, along with a strong positive correlation between $K_D$ and calculated temperatures in Udachnaya, it can be seen that samples with similar distribution coefficients (0.20–0.30 range) may have 100–200 °C differences in temperature (Fig. 5). Another important geochemical parameter, $X_{Mg}$, does not vary significantly in either analyzed (by Möss-
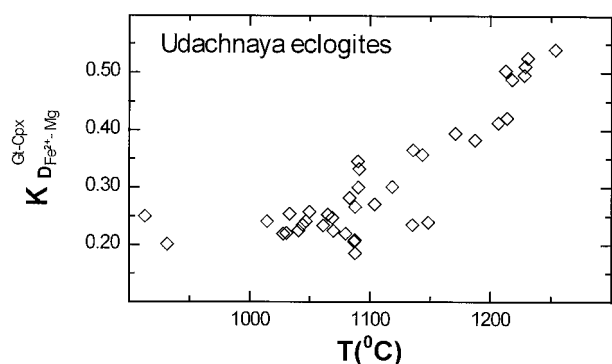
**FIGURE 5.** $K_D$ Fe²⁺-Mg distribution coefficient ($K_D$ = [Fe-$^{Cpx}$Mg$^{Gt}$]/[Fe$^{Gt}$Mg$^{Cpx}$]) vs. calculated temperature for Udachnaya eclogites (data from Sobolev et al. 1994).



**FIGURE 6.** Temperature of equilibration vs. Fe³⁺/ΣFe of garnet and clinopyroxene from eclogite U-107/2. Fe³⁺/ΣFe vs. temperature for garnet was calculated assuming constant Fe³⁺/ΣFe in clinopyroxene and vice versa.

bauer technique, Table 1) or tested (Table 2) samples with variation of Fe³⁺ contents. It can be seen in both Tables 1 and 2 that its variation does not exceed 2.5 relative percent.

Because a strong positive correlation exists between $K_D$ and calculated temperatures, one can expect a similar correlation between the latter and Fe³⁺ contents. Luth et al. (1990) stated that the amount of Fe³⁺ in garnets correlates with the pressure and temperature of the host xenolith. In general, Fe³⁺/ΣFe in garnet increases with increasing pressure and temperature. A positive correlation exists if different geothermometers (Wells 1977; Bertrand and Mercier 1985; Finnerty and Boyd 1987) and geobarometers (MacGregor 1974; Nickel and Green 1985) are applied. According to Luth et al. (1990), the amount of Fe³⁺ in garnets also correlates with differences in the textures of peridotites. Canil and O'Neill (1996) also showed that a positive correlation exists between Fe³⁺/ΣFe in garnet and temperature of equilibration. However, only a slight correlation was found between Fe³⁺ in clinopyroxene and temperature (Canil and O'Neill 1996).

We used the compositions of garnet and clinopyroxene from diamondiferous eclogite xenolith U-107/2 (data from Sobolev et al. 1994) to check the dependence of Fe³⁺ on temperature, using the geothermometer of Ellis and Green (1979). The Fe³⁺/ΣFe ratio in garnet was tested in the range from 8.0 to 70.0 at about seven different increments; each resulting composition was then used with coexisting clinopyroxene (same composition for each estimation). As shown in Figure 6, temperature increases with increasing Fe³⁺/ΣFe. The same calculation was performed with increasing Fe³⁺/ΣFe of clinopyroxene, which resulted in a negative correlation (Fig. 6). It is apparent that some compensation actually occurs with increasing $f_{O_2}$ of the system (as reflected in Fe³⁺ contents), thereby keeping temperatures from changing drastically. Note that this is true only when Fe²⁺-Mg exchange geothermometers are applied to the Gt-Cpx (e.g., eclogitic) mineral assemblage.

It is clear that more experimental work is needed on the Fe²⁺-Fe³⁺ substitution in eclogitic and peridotitic min-
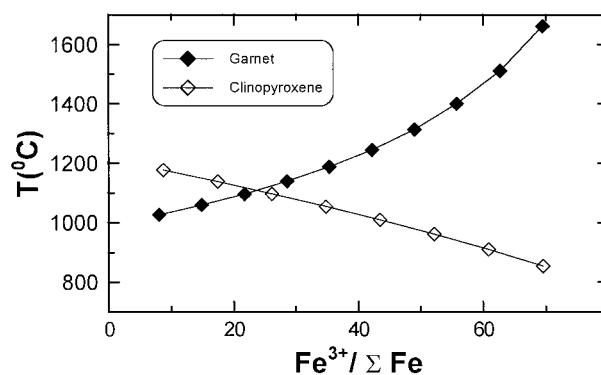
erals. Although uncertainties in the pressure and temperature estimates for mantle xenoliths do not affect the calculated $f_{O_2}$ significantly (Luth et al. 1990), the Fe³⁺ contents in mantle minerals provide information about the environment in which equilibration occurred. Due to the fact that all geothermometers for eclogites involve Fe²⁺-Mg exchange, Fe³⁺/ΣFe must be measured by a Mössbauer technique, and errors in temperature estimates for high-Na clinopyroxene eclogites using Fe³⁺/ΣFe from EMP analyses can seriously affect petrologic interpretation. Moreover, because a constant value of Mg²⁺ dilutes the effect of possible error in determination of Fe³⁺ contents, the uncertainty in this determination may affect most Fe²⁺ $K_D$ and Fe²⁺ diffusion coefficient values as well.

## SUMMARY

(1) The results of this study demonstrate that there is no direct (1:1) correlation between Mössbauer-measured and EMP-calculated values for Fe³⁺/ΣFe of eclogitic garnets and clinopyroxenes. Values of Fe³⁺/ΣFe calculated from EMP analyses of silicate minerals are generally inaccurate, although they do not greatly affect temperature estimates in eclogites due to compensation effects between garnet and clinopyroxene.

(2) In silicates, the precision in determination of $SiO_2$ largely controls the Fe³⁺/ΣFe values based on stoichiometry due to its abundance and +4 ionic charge contribution. In addition, high-Na clinopyroxenes (>4–5 wt% $Na_2O$) are even more sensitive to EMP error, and therefore can produce larger uncertainties in temperature estimates.

(3) The interpretation of the results of many previously published papers on Fe-Mg distribution between coexisting silicate minerals may be significantly changed due to possible error in Fe³⁺ estimation.

(4) Compensation effects between garnet and clinopyroxene exist, even with increasing $f_{O_2}$; this compensation moderates temperature from changing greatly with $K_D$ in eclogites.

(5) The uncertainty in $Fe^{3+}$ calculated from EMP analyses, due solely to inherent analytical precision, raises questions about most $Fe^{2+}$ $K_D$ and diffusion coefficient values determined with natural silicate minerals.

## ACKNOWLEDGMENTS

## REFERENCES CITED

Ai, Y. (1994) A revision of the garnet-clinopyroxene $Fe^{2+}$-Mg exchange geothermometer. Contributions to Mineralogy and Petrology, 115, 467–473.

Amthauer, G. (1976) Crystal chemistry of silicate garnets; Mossbauer and electronic absorption studies. Abstracts, International Geological Congress, 2, 14, 551–552.

Banno, S. (1970) Classification of eclogites in terms of physical conditions of their origin. Physics of the Earth and Planetary Interiors, 3, 405–421.

Bertrand, P. and Mercier, J-C.C. (1985) The mutual solubility of coexisting ortho- and clinopyroxene: towards an absolute geothermometer for the natural system? Earth and Planetary Science Letters, 76, 382–403.

Canil, D. and O'Neill, H.St.C. (1996) Distribution of ferric iron in some upper mantle assemblages. Journal of Petrology, 37, 609–635.

Coleman, R.G., Lee, D.E., Beatty, L.B., and Brannock, W.W. (1965) Eclogites and eclogites: their differences and similarities. Bulletin Geological Society of America, 76 p. 483–508.

Delaney, J.S., Bajt, S.R., Sutton, S.R., and Dyar, M.D. (1996) In situ microanalysis of $Fe^{3+}/\Sigma Fe$ in amphibole by X-ray absorption near edge structure (XANES) spectroscopy. In Mineral Spectroscopy: A Tribute to Roger G. Burns. The Geochemical Society Special Publication, 5, 165–171.

Dollase, W.A. and Newman, W.I. (1984) Statistically most probable stoichiometric formulae. American Mineralogist, 69, 553–556.

Dyar, M.D. (1984) Precision and interlaboratory reproducibility of measurements of the Mössbauer effect in minerals. American Mineralogist, 69, 1127–1144.

Ellis, D.J. and Green, D.H. (1979) An experimental study of the effect of Ca upon garnet-clinopyroxene exchange equilibria. Contributions to Mineralogy and Petrology, 71, 13–22.

Finger, L.W. (1972) The uncertainty in the calculated ferric iron content of a microprobe analysis. Year Book, Carnegie Institution of Washington, 71, 600–603.

Finnerty, A.A. and Boyd, F.R. (1987) Thermobarometry for garnet peridotites: basis for the of thermal and compositional structure of the upper mantle. In P.H. Nixon, Ed., Mantle xenoliths, p. 381–402. Wiley, New York.

Geiger, C.A, Armbruster, T., Lager, G.A., Jiang, K., Lottermoser, W., and Amthauer, G. (1992) A combined temperature dependent ⁵⁷Fe Moessbauer and single crystal X-ray diffraction study of synthetic almandine; evidence for the Gol'danskii-Karyagin effect. Physics and Chemistry of Minerals, 19, 2, 121–126.

Jacob, D., Jagoutz, E., Lowry, D., Mattey, D., and Kudrjavtseva, G. (1994) Diamondiferous eclogites from Siberia: Remnants of Archean oceanic crust. Geochimica et Cosmochimica Acta, 58, 5191–5207.

Jerde, E.A., Taylor, L.A., Crozaz, G., Sobolev, N.V., and Sobolev, V.N. (1993) Diamondiferous eclogites from Yakutia, Siberia: Evidence for a diversity of protoliths. Contributions to Mineralogy and Petrology, 114, 189–202.

Krogh, E.J. (1988) The garnet-clinopyroxene Fe-Mg geothermometer—a reinterpretation of existing experimental data. Contributions to Mineralogy and Petrology, 99, 44–48.

Luth, R.W. and Canil, D. (1993) Ferric iron in mantle-derived pyroxenes and a new oxybarometer for the mantle. Contributions to Mineralogy and Petrology, 113, 2, 236–248.

Luth, R.W., Virgo, D., Boyd, F.R., and Wood, B.J. (1990) Ferric iron in mantle-derived garnets. Implications for thermobarometry and for the oxidation state of the mantle. Contributions to Mineralogy and Petrology, 104, 56–72.

MacGregor, I.D. (1974) The system $MgO-Al_2O_3-SiO_2$: solubility of $Al_2O_3$ in enstatite for spinel and garnet peridotite compositions. American Mineralogist, 61, 603–615.

McCammon, C.A. (1994) A Mössbauer milliprobe: Practical considerations. Hyperfine Interactions, 92, 1235–1239.

McCammon, C.A., Chaskar, V., and Richards, G.G. (1991) A technique for spatially Mössbauer spectroscopy applied to quenched metallurgical slags. Measurement Science and Technology, 2, 657–662.

McCammon, C.A., Chinn, I.L., Gurney, J.J., and McCallum, M.E. (1995) Determination of the ferric iron content of diamond inclusion from George Creek using Mössbauer spectroscopy with implications for mantle oxygen fugacity during diamond genesis, Extended Abstracts., 6th International Kimberlite Conference, 359–361.

McCammon, C.A., Chinn, I.L., Gurney, J.J., McCallum, M.E. (1998) Ferric iron content of mineral inclusions in diamonds from George Creek, USA determined using Mössbauer spectroscopy: Implications for oxygen fugacity. Contribution to Mineralogy and Petrology, in press.

Nickel, K.G. and Green, D.H. (1985) Empirical geothermobarometry for garnet peridotites and implications for the nature of the lithosphere, kimberlites and diamonds, Earth and Planetary Science Letters, 73, 158–170.

O'Neill, H.St.C. and Wood, B.J. (1979) An experimental study of Fe-Mg partitioning between garnet and olivine and its calibration as a geothermometer. Contributions to Mineralogy and Petrology, 70, 59–70.

Powell, R. (1985) Regression diagnostics and robust regression in geothermometer/geobarometer calibration: the garnet-clinopyroxene geothermometer revisited. Journal of Metamorphic Geology, 6, 231–243.

Råheim, A. and Green, D.H. (1974) Experimental determination of the temperature and pressure dependence of the Fe-Mg partition coefficient for coexisting garnet and clinopyroxene. Contributions to mineralogy and Petrology, 48, 179–203.

Snyder, G.A., Taylor, L.A., Crozaz, G., Halliday, A.N., Beard, B.L., Sobolev, V.N., and Sobolev, N.V. (1997) The origins of Yakutian eclogite xenoliths. Journal of Petrology, 38, 85–113.

Sobolev, N.V. (1964) Paragenetic types of garnets, Nauka Publishing House, Moscow (in Russian).

———(1974) Deep-seated inclusions in kimberlites and the problem of the composition of the upper mantle, Nauka Publishing House, Novosibirsk, 264 p. (in Russian). English translation published by American Geophysical Union, Washington, D.C., 1977, 279 p.

Sobolev, V.N., Taylor, L.A., Snyder, G.A., and Sobolev, N.V. (1994) Diamondiferous eclogites from the Udachnaya kimberlite pipe, Yakutia, Siberia. International Geology Review, 36, 42–64.

Sobolev, V.N., Taylor, L.A., Snyder, G.A., Sobolev, N.V., Pokhilenko, N.P., and Kharkiv, A.D. (1997) A unique metasomatised peridotite from the Mir kimberlite, Siberian Platform. Proceedings of the 6ᵗʰ International Kimberlite Conference, Novosibirsk, Russia 1997, Geology and Geophysics, 38, 206–215.

Taylor, L.A. and Neal, C.R. (1989) Eclogites with oceanic crustal and mantle signatures from the Bellsbank kimberlite, South Africa, part I: Mineralogy, petrography, and whole rock chemistry. Journal of Geology, 97, 551–567.

Wells, P.R.A. (1977) Pyroxene thermometry in simple and complex systems. Contributions to mineralogy and Petrology, 62, 129–139.

Wood, B.J. and Virgo, D. (1989) Upper mantle oxidation state; ferric iron contents of lherzolite spinels by (super 57) Fe Moessbauer spectroscopy and resultant oxidation fugacities. Geochimica et Cosmochimica Acta, 53, 1277–1291.