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Comment on “Mantle eclogites: evidence of igneous fractionation in the mantle” by J.R. Smyth, F.A. Caporuscio, and T.C. McCormick

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1. Introduction

Smyth et al. [1] have recently reported major and trace element mineral chemistry and petrography of a suite of eclogite xenoliths from the Bellsbank kimberlite, South Africa, which they use to support an igneous fractionation model for eclogite petrogenesis [2–7]. Paramount to this model are the presence of garnet and kyanite exsolution textures from clinopyroxene, and the continuum of compositions presented on an ACF diagram. But, there is another currently popular model for the petrogenesis of eclogite xenoliths in kimberlite as the products of subducted oceanic crust and transformation to eclogite (e.g., [8–15]). To date, there has been no really definitive test or analysis that can be used to verify or negate either model. However, isotopic analyses of the “ultrapure” (involving acid leaching; [14,15]), primary eclogite phases may hold the key.

We have also analyzed a suite of eclogite xenoliths from this same Bellsbank kimberlite [12,14,15] and observed three different eclogite suites, which we have divided into Groups A, B, and C on the basis of mineral, REE, and isotope chemistry. The pertinent criteria used in these divisions are summarized in Table 1. Note that our Group C eclogites correspond to the grosspydites and kyanite eclogites of Smyth et al. [1]. Although we have analyzed a smaller number of eclogites for mineral chemistry than Smyth et al. [1], we have performed detailed trace element and very importantly, isotopic analyses on “ultrapure” mineral

separates from a greater number of samples and have thereby extracted considerably more information. (We emphasize the use of “ultrapure” mineral separates here since the use of simple binocular microscopic picking, such as used by Smyth et al. [1], is not sufficient to give accurate trace and isotopic elemental analyses.) With a greater number of samples, the boundaries of these groups, as defined by mineral chemistry alone, expand and may give the impression of an apparent continuum, consistent with igneous fractionation. Furthermore, we agree that the garnet and kyanite exsolution from clinopyroxene described by Smyth et al. [1] from Bellsbank grosspydites probably does represent cooling from high temperatures at pressures > 30 kbar.

The fact that the isotopic signatures of Group B and C eclogites are consistent with an oceanic crustal protolith does not preclude various degrees of igneous activity. This includes small amounts of partial melting, which probably occurred, all the way to complete melting, which probably did not. In fact, it is naive to think that a crustal protolith subducted into a high pressure and temperature regime would not experience some degree of partial melting. As such, we would not expect bulk compositions to exactly conform with those of oceanic crust seen present day. Indeed, such depletions are witnessed in the Nd isotope data, where Sm/Nd ratios are as high as 1.615 in FRB438-7 relative to a MORB Sm/Nd ratio of ≈ 0.3 .

That these eclogites have witnessed at least one

TABLE 1

Compositional ranges of the three eclogite groups found at the Bellsbank kimberlite

	Group A Mantle cumulate	Group B Subducted oceanic crust MORB	Group C Cumulate
Clinopyroxene			
Na ₂ O	0.6–2.8	3.7–5.7	7.6–8.7
Al ₂ O ₃	0.9–3.5	5.5–8.6	14.8–17.0
Cr ₂ O ₃	0.1–1.3	< 0.1	< 0.1
Garnet			
Comps.	Mg-rich	Fe-rich	Ca-rich
Cr ₂ O ₃	1.0–2.2	< 0.1	< 0.1
Mineral Mg#’s			
Garnet	78–93	57–59	67–73
Cpx	90–96	86–89	90–93
Opx	89–93	–	–
Whole-rock Mg#	82–89	61–67	72–78
REE’s	LREE-enriched HREE-depleted	LREE-depleted HREE-enriched	LREE-depleted HREE-depleted
Eu anomaly	No	No	Yes
Isotopic comps.			
$\delta^{18}\text{O}$	+5.1 to +5.6	+2.9 to +4.0	+3.4 to +4.7
$\epsilon_{\text{Nd}}^{143}$	-19 to -16	+39 to +241	+46 to +112
$^{87}\text{Sr}/^{86}\text{Sr}$	0.7042–0.7046	0.7086–0.7100	0.7083–0.7101

depletion is supported by considering the age of these xenoliths. Previously reported Nd isotope data for eclogite xenoliths suggest an Archean age (e.g., [10,11]) of crystallization or recrystallization. Sleep and Windley [16] proposed that the Archean mantle was much hotter than present day. Therefore, subducted oceanic crust (involving transformation to eclogite) introduced into this hot mantle will be heated to high temperatures and experience a change in bulk composition through depletion(s). Post-Archean mantle cooling will facilitate the exsolution textures in the eclogitic residue. Such a scenario can account for the exsolution textures in the Bellsbank grosspyrites. Therefore, because of the various depletions that these eclogites have experienced, it is *not the bulk major element compositions which are definitive*. Rather, it is the trace element and isotopic signatures which should be considered in the formulation of a petrogenetic model for these eclogites.

On the basis of trace element and Sr, Nd, and O isotopic signatures, we have concluded that our Group A eclogites are high pressure (mantle) igneous cumulates, but that the Group B and C eclogites represent ancient, subducted oceanic crust. Group B eclogites represent the basaltic portion

and Group C the plagioclase-rich cumulates. It is the purpose of this contribution to briefly outline evidence which negates a cumulate origin for most (i.e., $\approx 80\%$) of the Bellsbank eclogite xenoliths (i.e., Groups B and C), including those reported by Smyth et al. [1].

2. Evidence against high-pressure igneous fractionation

2.1. Whole-rock compositions

Whole-rock compositions have been reconstructed using the analyses of ultrapure garnet and clinopyroxene mineral separates and modes to remove the effects of secondary metasomatism and kimberlite infiltration. If igneous fractionation of either garnet and clinopyroxene or a hyperaluminous clinopyroxene produced the range of Bellsbank eclogite compositions, then elements compatible with these minerals should form positive correlations with Mg#. For basaltic systems, partition coefficients for Co of up to 1.9 in clinopyroxene [17] and up to 1.0 in garnet [18] have been reported. In the absence of sulfide minerals such as we have in these eclogites, Co should act

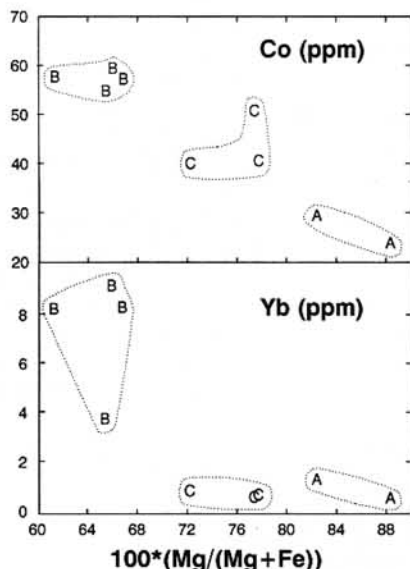


Fig. 1. Reconstructed whole-rock plots of $Mg\#$ vs Co (ppm) and Yb (ppm) for the three Bellsbank eclogite groups.

as a lithophile element. For Yb, partition coefficients of up to 1.3 have been reported for clinopyroxene [19] and up to 8.0 for garnet [18]. Abundances of Co and Yb, both of which should be concentrated in the fractionating minerals, have been plotted against $Mg\#$ (Fig. 1). Groups A, B, and C eclogites cannot be related by igneous fractionation because negative, rather than positive correlations are displayed in Fig. 1—the opposite of what is required. Furthermore, Smyth et al. [1] suggest that the grosspydites (our Group C's) were an early fractionate. However, although they possess high whole-rock $Mg\#$'s, they are intermediate relative to the whole-rock $Mg\#$'s of Groups A and B (Table 1), and also contain the highest CaO and Al_2O_3 contents of all the eclogites analyzed.

2.2. REE compositions

Smyth et al. [1] infer that Bellsbank grosspydites (our Group C) are produced by high-pressure igneous fractionation and accumulation of hyper-aluminous clinopyroxene, because of garnet (and kyanite) exsolving from clinopyroxene. However, we also have found garnet exsolution from clinopyroxene in the bi-mineralic Groups A and B

Bellsbank eclogites, which should also have formed from accumulation of hyper-aluminous clinopyroxene, if the model of Smyth et al. [1] is correct. However, these authors call upon clinopyroxene, rather than plagioclase accumulation, to explain the positive Eu anomaly present in the Bellsbank grosspydites/Group C eclogites (Fig. 2). If such hyper-aluminous clinopyroxene accumulation generates the positive Eu anomaly observed in these eclogites, why do Group A and B eclogites (Table 1; Fig. 2) possess no anomaly? In these eclogites, garnet exsolution is also present and, by inference, should also have been produced by clinopyroxene accumulation? It is this observation which we consider is evidence of plagioclase involvement (accumulation) during the petrogenesis of these eclogites.

Group C eclogites would be expected to have an inherent LREE-enriched REE profile if isochemical eclogite transformation occurred and no subsequent depletions were experienced, because the inferred protolith is considered to be a plagioclase-rich cumulate. However, the fact that we have consistent depletion of the LREE over the HREE indicates that these eclogites have experienced at least one depletion. It is the presence of the positive Eu anomaly in these Group C eclogites that suggests plagioclase accumulation. Note that *the REE profiles of the Group C eclogites do not exhibit a convex-upward pattern* (depletion of the LREE and HREE over the MREE, Fig. 2),

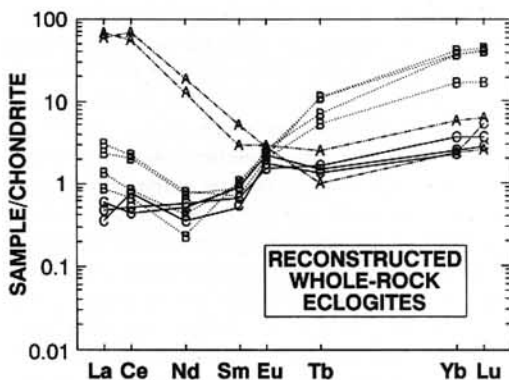


Fig. 2. Reconstructed whole-rock REE profiles of Group A, B, and C Bellsbank eclogites. These were constructed using INA and modal analyses from Taylor and Neal [12]. Note that only Group C eclogites possess a positive Eu anomaly.

TABLE 2
Reconstructed whole-rock isotope data

	Rb (ppm)	Sr (ppm)	$^{87}\text{Rb}/^{86}\text{Sr}$ (ppm)	$^{87}\text{Sr}/^{86}\text{Sr}$ (p)	$^{87}\text{Sr}/^{86}\text{Sr}$ (120)	$\delta^{18}\text{O}$ (‰)	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$ (p)	$^{143}\text{Nd}/^{144}\text{Nd}$ (p)	$^{143}\text{Nd}/^{144}\text{Nd}$ (120)	ϵ_{Nd}
<i>Group A</i>												
437-1 ^b	0.011	228	0.00014	0.70420	0.70420	+5.6	0.91	11.6	0.04762	0.511801	0.511676	-16
437-2	0.252	185	0.00394	0.70455	0.70455	+5.1	0.58	10.1	0.03460	0.511682	0.511591	-19
<i>Group B</i>												
438-2 ^b	0.002	9.04	0.00064	0.70936	0.70935	+3.3	0.21	0.24	0.53016	0.518848	0.517459	+121
438-5	0.162	11.6	0.04041	0.70924	0.70901	+4.0	0.20	0.44	0.27519	0.515531	0.514810	+56
438-7 ^b	0.0005	6.36	0.00023	0.71002	0.71002	+3.0 ^a	0.21	0.13	0.98017	0.524993	0.522426	+241
2791-21	0.174	8.43	0.05972	0.70874	0.70864	+2.9	0.13	0.46	0.17106	0.514659	0.514211	+39
<i>Group C</i>												
437-5	0.068	4.36	0.04513	0.70857	0.70832	+3.4	0.11	0.22	0.31118	0.518360	0.517545	+112
437-7	0.071	4.02	0.05111	0.71027	0.70998	+4.4	0.13	0.36	0.21486	0.515017	0.514454	+46
2791-34	0.141	8.77	0.04653	0.71032	0.71006	+4.7	0.17	0.30	0.34317	0.516764	0.515865	+80

^a Assumes S^{18}O (Cpx) = S^{18}O (Qtz).

^b Data from Shervais et al. [9].

Values were calculated from the mineral data and modal analyses reported in Taylor and Neal [12]. Present day $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios are corrected to the age of eruption of the Bellsbank kimberlite (120 Ma, [29]).

which Smith et al. [1] argue is indicative of hyperaluminous clinopyroxene accumulation. We conclude that plagioclase accumulation in the oceanic protolith prior to eclogite transformation is responsible for generating the positive Eu anomaly in Group C eclogites (Fig. 2).

2.3. Sr, Nd and O isotopes

The range of Sr, Nd, and O isotopes (Table 2) from the three Bellsbank eclogite groups is a compelling argument against high pressure (mantle) igneous fractionation for the petrogenesis of all three eclogite groups. Through REE and Sr and Nd isotope modeling, we have demonstrated that the Group A eclogites were formed as igneous cumulates from a Group II kimberlitic magma at ≈ 1 Ga [14,15]. The range of $\delta^{18}\text{O}$ compositions of the Group A eclogites (+5.1 to +5.6) is consistent with a mantle origin [20–22]. However, such modeling cannot account for the isotopic compositions of the Group B and C eclogites (e.g., $\delta^{18}\text{O} = +2.9$ to +4.7).

The oxygen isotopic signatures of the Group B ($\delta^{18}\text{O} = +2.9$ to +4.0) and Group C ($\delta^{18}\text{O} = +3.4$ to +4.7) eclogites demonstrate that the ultimate origin for these xenoliths could not have been in a mantle regime [20–22]. Smyth [23] has suggested that oxygen isotope fractionation between minerals could occur on the basis of an electrostatic site potential model. Therefore, disequilibrium between garnet and clinopyroxene pairs, used in the calculation of our whole-rock eclogite composition, should be evident. As can be seen in Fig. 3, clinopyroxene and garnet from all but two Group C eclogites (FRB437-5 and PHN2791-34) are in oxygen isotope equilibrium. The garnet from one Group B eclogite yielded a $\delta^{18}\text{O}$ with large errors associated with it due to lack of sample. Eclogites FRB437-5 and PHN2791-34 exhibit signs of disequilibrium and yield $\Delta_{\text{cpx-gt}}$ values of +0.8 and -0.4, respectively. This disequilibrium certainly is not enough to yield a whole-rock $\delta^{18}\text{O}$ which is within mantle values. Furthermore, the general oxygen isotope equilibrium exhibited by the garnet-clinopyroxene pairs demonstrates that the low $\delta^{18}\text{O}$ values in Group B and C eclogites are not a product of electrostatic site potential differences between constituent minerals.

Smyth et al. [1] acknowledge the problem of low $\delta^{18}\text{O}$ values in Group B and C eclogites,

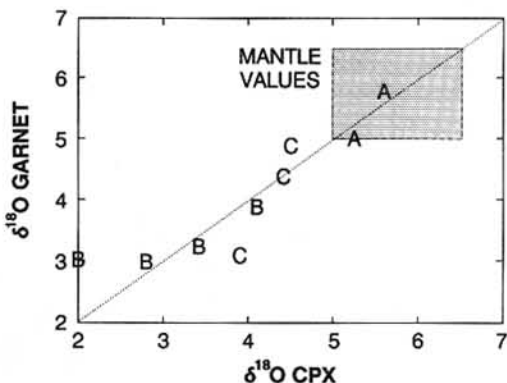


Fig. 3. $\delta^{18}\text{O}$ Cpx versus $\delta^{18}\text{O}$ garnet to demonstrate that disequilibrium between garnet and clinopyroxene mineral pairs cannot account for the range in $\delta^{18}\text{O}$ values between all three eclogite groups.

suggesting the spread in $\delta^{18}\text{O}$ values may be due to "low-temperature metasomatic processes within the mantle or post-emplacment alteration". The purity of our mineral separates precludes any contamination either by metasomatic or post-emplacment alteration products. Also, mantle temperatures are too high to allow for fractionation of $\delta^{18}\text{O}$ from typical mantle values of 5–6‰ to 3‰ (Table 2). In the final sentence of their contribution, Smyth et al. [1] suggested that such low $\delta^{18}\text{O}$ signatures can only be generated in a crustal environment, "possibly through low-temperature alteration and subsequent subduction and melting of oceanic crust". We feel that in these samples, this is the *only* way of generating such depleted oxygen isotopic signatures.

The range in Sr and Nd isotopic signatures also argue against a petrogenesis by high-pressure (mantle) igneous fractionation for the Group B and C eclogites. Smyth et al. [1] have chosen to ignore the already published isotope results for Bellsbank eclogites (e.g., [9,14]) in the formulation of their igneous fractionation model. Furthermore, Smyth et al. [1] misquoted the literature when they state that Shervais et al. [9] explained the unusual Nd isotopes (i.e., $\epsilon_{\text{Nd}} > +200$) by metasomatic alteration at low pressures; this process, in the form of seawater alteration or exchange with a high Rb reservoir, was inferred to account for the radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ composition of these eclogites [9].

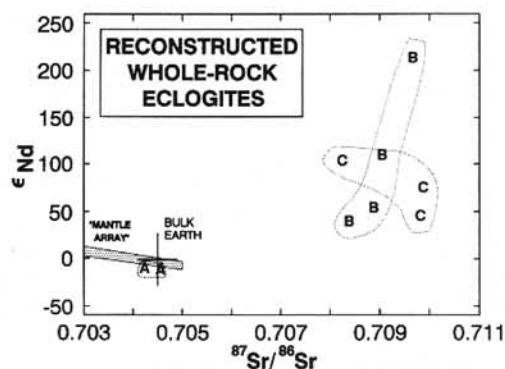


Fig. 4. ϵ_{Nd} versus $^{87}Sr/^{86}Sr$ for reconstructed whole-rock eclogites (data from Neal et al. [15]). Isotopic ratios have been normalized to the age of eruption of the Bellsbank kimberlite (120 Ma).

The Sr and Nd isotopic signatures of Group B and C eclogites are totally different from those of the Group A eclogites (Table 2 and Fig. 4), which we have demonstrated are mantle cumulates [12,14,15]. The range of Nd isotopes from the Group B and C eclogites can be explained by simple two-component mixing between a depleted eclogite end-member and the Bellsbank kimberlite [15]. This may be envisaged as cryptic metasomatism and can be witnessed in the Sm/Nd ratios (Fig. 2 and Table 2). The depletion of Nd isotopes in the eclogite end-member is extreme ($\epsilon_{Nd} = +241$; Sm/Nd = 1.615). Such a Nd isotopic signature could be generated by ancient high-pressure igneous fractionation of clinopyroxene and garnet. In fact, Nd model ages suggest final equilibration at T_{CHUR} of ≈ 2.4 Ga. However, the radiogenic $^{87}Sr/^{86}Sr$ ratios of these eclogites demonstrate that the Sr and Nd isotopes are decoupled (Table 1; Fig. 3). No igneous fractionation process will preferentially deplete the cumulate in Nd and Sr (thus increasing both Sm/Nd and Rb/Sr ratios), especially if clinopyroxene and garnet are liquidus phases. Note that in Table 2, it is evident that the Rb contents of these eclogites are extremely low, but that Sr contents are relatively high. As such, the $^{87}Sr/^{86}Sr$ has not been generated by in-situ decay of ^{87}Rb . Rather, the Sr isotope signatures are inherited at the time of eclogite formation. Therefore, we conclude that the range in Sr, Nd, and O isotopic signatures of the group B and C eclogites cannot be readily

generated by a closed-system high-pressure or mantle igneous fractionation process. Indeed, even an open-system fractionation process (i.e., AFC) would have trouble in generating the observed ranges in isotopic ratios, especially the high $^{87}Sr/^{86}Sr$ because of the unusual composition of the required assimilate.

3. A model for Bellsbank eclogite petrogenesis

Radiogenic isotope studies of eclogite xenoliths in kimberlite xenoliths suggest final equilibration during the Archean or early Proterozoic (e.g., [9–11,14,15]). Because of their antiquity, degree of secondary alteration, and the depletions experienced, the bulk major-element compositions will be meaningless in trying to define a petrogenetic model for these eclogites. Therefore, we must rely upon distinctive geochemical signatures to unravel the origin and evolution of such xenoliths.

The mineral Kd's coupled with trace element and isotopic signatures of the Group A eclogites are consistent with a petrogenesis as a mantle cumulate [14,15] from a Group II kimberlitic magma. Cooling of these cumulates would produce the observed exsolution garnet-clinopyroxene features observed in the Group A eclogites. This kimberlitic magma was unrelated to the Bellsbank kimberlite.

The depleted oxygen isotopic signatures of the Group B and C eclogites (Table 1) are consistent with hydrothermal alteration of oceanic crust (e.g., [24,25]). The range in $\delta^{18}O$ observed in these eclogites is not a function of disequilibrium between garnet and clinopyroxene (Fig. 3) caused by different electrostatic site potentials for oxygen in these minerals. Seawater alteration can also account for the radiogenic $^{87}Sr/^{86}Sr$ ratios of these eclogites, as the Sr isotopic signature of Precambrian seawater is not well defined [26]. Furthermore, seawater-altered basalts may possess $^{87}Sr/^{86}Sr$ ratios which are greater than that of seawater [27,28]. The seawater-alteration process does not radically affect the Nd isotopes. Generation of the high Sm/Nd ratio observed in the depleted eclogite end-member occurs during open-system eclogite transformation: garnet and clinopyroxene are inherently highly LREE-depleted (> MORB), requiring a loss of the LREE during this process. As stated above, introduction

of this subducted slab into the hot mantle (c.f., Sleep and Windley [16]) may have been responsible for not only raising the eclogites to high temperatures, but resulted in generating the depleted Sm/Nd ratio by small degrees of partial melting of the subducted slab. Subsequent cooling of the mantle to present-day geothermal gradients will produce the observed exsolution features observed in both B and C Bellsbank eclogite groups.

4. Conclusions

Smyth et al. [1] has used primarily petrography and mineral chemistry as evidence for igneous fractionation within the mantle for Group C eclogites. When taken in isolation, such conclusions may be drawn from this evidence. However, *it is the range in trace element and Sr, Nd, and O isotopic signatures which cannot be reconciled with an igneous fractionation petrogenesis.* Indeed, these eclogites, along with our Group B's, possess REE and isotopic signatures which are not consistent with a mantle petrogenesis. Granted they have experienced extreme metamorphism (i.e., high temperatures and probable low degrees of partial melting, followed by cooling and garnet exsolution from clinopyroxene). However, the protoliths were of oceanic crustal origin and were emplaced into the mantle during late Archean/early Proterozoic times. Because of their complicated history, we do not expect the major-element bulk compositions of these eclogites to reflect the crustal progenitor, and we must rely upon characteristic *signatures* to unravel the petrogenesis of these eclogites. Failure to account for the REE and isotopic signatures of such xenoliths (obtained by analysis of *ultrapure* garnet and clinopyroxene mineral separates) can result in misleading petrogenetic interpretations.

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