"Majorite" and "Silicate Perovskite" Mineral Compositions in Xenoliths from Malaita

The report by Collerson *et al*. (*1*) of transition zone and lower mantle phases in mantle-derived material from alnöite pipes on Malaita, Solomon Islands, is highly significant in view of the rarity of such natural assemblages. Malaita is part of the obducted southern margin of the Ontong Java Plateau, an area of over-thickened oceanic crust (*2–7*) that was emplaced by two major magmatic events at ~122 Ma and ~90 Ma (*8*, *9*) and two minor events at ~62 Ma and ~36 Ma (*10*, *11*). Recent seismic evidence has demonstrated the presence of a "mantle root" beneath the Ontong Java Plateau, interpreted to represent the remains of a plume head (*7*, *12*), that has been repeatedly tapped (*10*, *11*). The alnöite magmas were emplaced in the OJP at ~34 Ma (*13*).

Collerson *et al*. (*1*) used mineral chemistry, petrography, and infrared spectral data to identify majorite, silicate perovskite, and diamond in mantle xenoliths entrained by alnöite magmas from eastern Malaita. We urge extreme caution in interpreting those data in terms of ultrahigh pressure mineral phases, especially because definitive x-ray data are lacking. Mineral chemistry (Fig. 1 and Tables 1 to 3) previously reported for mantle xenoliths from the Malaitan alnöites (*14–20*) shows compositions akin to those reported by Collerson *et al*. (*1*); the strong compositional similarity between the two data sets suggests that the minerals have a common origin. The question is, is that origin deep in the mantle, or shallow?

Mineral chemistry. It is not possible to definitively identify majorite on the basis of mineral chemistry alone (*21–24*). Diagrams of the type depicted here (Fig. 1) and by Collerson *et al*. [figure 2 of (*1*)] do not show direct evidence of Si in octahedral sites, although it can be inferred. The covariation between Si and Al + Cr can be interpreted either as majorite substitution in a garnet structure (Si-Al substitution in the octahedral site) or tschermak substitution in a pyroxene structure (Si-Al substitution in the tetrahedral site). The distinction between majorite (cubic) and pyroxene is best made by x-ray diffraction, but such data were not included in (*1*). The compositions defined as majorite by Collerson *et al*. (*1*) are consistent with pyroxene containing up to 30 mole percent tschermak ($R^{2+}AlAlSiO_6$) in solid solution with either diopside or enstatite. Substitution is 2Al for R^{2+} + Si, for which the Malaita "majorites" and pyroxenes show an excellent linear correlation (Fig. 1).

Practically all mineral compositions (and, as discussed below, textures as well) reported by Collerson *et al*. (*1*) as being of ultrahigh-pressure origin are also found in megacrysts and spinel and garnet-spinel peridotite xenoliths from the Malaitan alnöites (Tables 1 to 3). For example, the garnets postulated in (*1*) to have equilibrated at >6 GPa (equivalent to ~200 km depth) have essentially the same compositions as samples reported by other workers (*14–20*) as equilibrating at <3.6 GPa, or ~120 km (Table 1). We believe that the phases described as majorite and perovskite in (*1*) are actually pyroxenes and amphiboles (Tables 1 to 3). For example, the E-type "majorites" of (*1*) have a composition similar to that of the bronzite megacrysts reported by Nixon and Boyd (*15*) and of an amphibole inclusion in a garnet megacryst (Table 2). The P-type majorite compositions (Table 2) are similar to amphibole and Al-rich clinopyroxene of metasomatic origin (*19*). The E-type perovskite compositions (*1*) are remarkably similar to primary orthopyroxenes from both garnet-spinel and spinel peridotites (Mg-perovskite; Table 3) and clinopyroxene inclusions in garnet megacrysts (Ca-perovskite; Table 3). The P-type Ca-perovskites identified in (*1*) are compositionally similar to secondary, retrograde clinopyroxene found between garnet and spinel (*16*, *17*). It is evident from photomicrographs (Fig. 2, A and B, that the secondary clinopyroxene and amphibole found between garnet and spinel are in a state of disequilibrium with the primary peridotite minerals (Fig. 2, A and B; *16–18*).

Ultra-high-pressure "majorite" [analyses 161 and 159, table 2 of (*1*); analysis KC-98-16, table 3 of (*1*)] are clearly deficient in Al + Cr (1.85 to 1.55). According to (*1*), such a deficiency is an evidence for ultradeep origin. Actually, these three analyses have more than eight cations for 12 oxygen—that is, they are Fe^{3+}-rich, which explains the Al + Cr deficiency in the octahedral sites. A simple Fe^{2+}/Fe^{3+} calculation allows the (Al + Cr + Fe^{3+}) site to be filled to two cations per 12 oxygen (or one cation per six oxygens for pyroxenes). Moreover, the comparison of KC-98-16 with analysis 23 from ultrahigh pressure experiments of (*25*) at 2.3 GPa is disputable. The two analyses differ from each other with respect to their Al/Si ratio, Mg mole fraction, and calcium, which is three times higher in the Malaita analyses of (*1*). Taking into account precision now attainable by modern microprobe, an oxide total of 97.08 would generally be rejected, unless there were reasons to assume that Fe^{3+} or that the mineral was hydrous, as in amphibole.

Collerson *et al*. (*1*) note that high-Al orthopyroxene is found only in reaction products produced via the breakdown of olivine and garnet at temperatures greater than 1500°C (*26*, *27*). The upper mantle temperature, based upon geothermometry (*14–18*), is ~900 to 1250°C. The plume head responsible for the generation of the Ontong Java Plateau, however, was likely hotter than ambient upper mantle by up to 400°C (*28–31*).
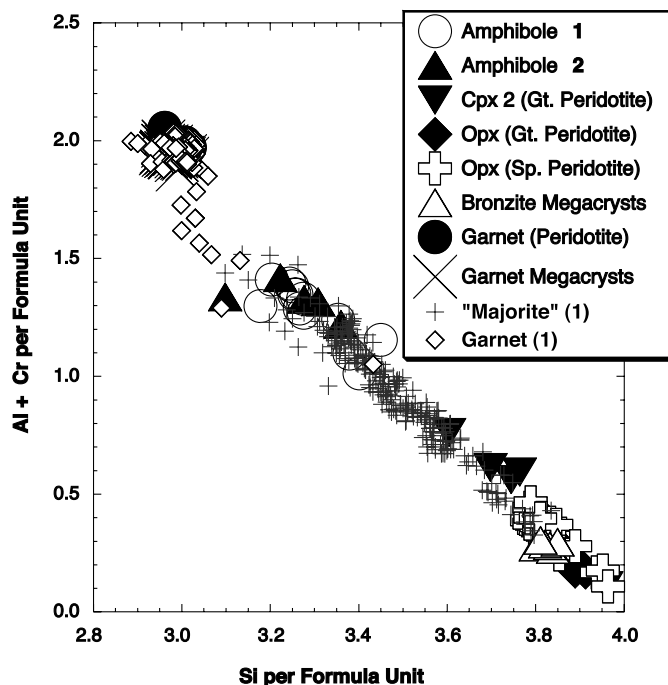

Fig. 1. Comparison of mineral compositions reported by Collerson *et al*. (*1*) with those previously reported for the xenolith suite from the Malaitan alnöites (*14–20*). The compositions interpreted as majorite (*1*) are the same as those interpreted as clinopyroxene, orthopyroxene, and amphibole (*14–20*). All data have been recalculated on the basis of 12 oxygen atoms.

Such a temperature would allow the development of Mg-pyroxenes with a large tschermak component at least below ~100 km. The recognition of a low-velocity root or remnant plume head ~200 km across and ~300 km deep (7, 12) provides a hint of the effect that the surfacing of the hot Ontong Java Plateau plume likely had upon the mantle sampled by the alnöitic magmas. Therefore, we suggest that Collerson et al. (1) may have found xenoliths that equilibrated deeper than 120 km, but not at the transition zone or in the lower mantle.

Furthermore, those xenoliths probably equilibrated above the diamond stability field, even though Collerson et al. (1) reported the presence of microdiamond in

their sample suite. The occurrence of a ~60-μm-diameter microdiamond inclusion in an E-type pyrope is the first report of diamond in a garnet mantle xenolith [see discussion in (32)] and is the only suggestion of a "deep" origin. However, the identification of the microdiamond is based upon what is not present. A negative result based on SiO_2, MgO, FeO, and CaO does not preclude a plethora of other minerals, such as titanites, sulfates, phosphates, platinum-group element minerals, and Ni-Cu sulfides (33–35). Furthermore, Collerson et al. [caption to figure 4B of (1)] state that Raman spectroscopy has "confirmed that this phase is carbon." In view of the power of Raman spectroscopy and the extensive literature on carbon species, it is surprising that the positions of the

Raman peaks were not reported. Without these and, possibly, x-ray data, the existence of diamond in these xenoliths remains equivocal. An alternative interpretation of figure 4A of (1) is that it represents a hole in the slide that trapped carbon from the coating for microbeam analysis and that it was this carbon that was identified by Raman spectroscopy. If diamond is present, then the possibility that this diamond was introduced during the cutting and polishing of the sample needs to be fully discounted. In essence, greater detail and more information are required to establish the existence of diamond in the Malaitan alnöite xenolith suite.

Geothermobarometry. Previously, geothermobarometric calculations using garnet peridotite, garnet-spinel peridotite, and megacryst

Table 1. Comparison of garnet and spinel compositions between those reported for the xenolith suite from Malaita (14–20) and those of Collerson et al. (1).

Mineral:	Garnet												Spinel			
Rock (ref.):	E type (1)				Megacryst (14–20)			P type (1)		Peridotite (14–20)			(1)			(14–20)
Sample #:	KC 99M3 #62	KC 26 #18	#53	#54	PHN 3882	PHN 3943	CRN 61	KC 98A #12	KC 98-16 #31	CRN 209	CRN 213		#51	#160	KC 98B	CRN 60*
SiO_2	41.73	41.48	41.18	41.74	41.41	42.00	41.06	41.87	42.26	41.67	41.56		0.23	0.29	0.09	0.00
TiO_2	0.44	0.70	0.56	0.56	0.57	0.57	0.65	0.07	0.11	0.56	0.08		0.37	0.27	0.11	0.39
Al_2O_3	23.18	22.71	23.18	23.05	23.20	23.11	23.16	21.80	22.38	20.06	24.00		59.87	58.20	60.56	59.88
Cr_2O_3	0.09	0.03	0.05	0.03	0.06	0.07	0.09	2.65	1.59	4.90	0.68		0.13	0.17	0.76	0.59
FeO	11.61	11.20	11.55	11.81	11.14	11.07	11.41	6.65	8.12	6.23	8.78		24.60	25.77	20.68	19.20
MnO	0.32	0.30	0.36	0.23	0.30	0.30	0.32	0.36	0.38	0.24	0.47		0.24	0.32	0.30	0.27
MgO	18.05	17.83	18.44	18.16	19.27	19.52	18.27	19.61	20.35	20.86	19.46		16.13	15.77	17.63	19.69
CaO	4.03	4.71	4.33	3.72	4.78	4.83	4.63	5.85	3.71	5.54	4.79		0.01	0.00	0.00	0.00
Total	99.45	98.96	99.65	99.30	100.73	101.47	99.59	98.88	98.90	100.06	99.82		101.58	100.79	100.13	100.02
$Mg/(Mg+Fe^{2+})$	0.735	0.739	0.740	0.733	0.755	0.759	0.740	0.840	0.817	0.856	0.798		0.539	0.522	0.603	0.646
$Ca/(Ca+Mg)$	0.138	0.160	0.144	0.128	0.151	0.151	0.154	0.177	0.116	0.160	0.150					

*Inclusion in a garnet megacryst.

Table 2. Comparison of "majorite" mineral compositions (1) with compositions of minerals from the xenolith suite of the Malaitan alnöites previously reported in (14–20).

Mineral:	"Majorite"							"UHP majorite"					
Rock (ref.):	E type (1)			Mega. (14–20)	P type (1)		Perid. (14–20)	E type (1)		Mega. (14–20)	P type (1)	Perid. (14–20)	
Sample #:	KC 98-18 #78	KC 99M5 #240	#56	PHN 3556/3A	KC 98-16 #18	KC 98-16 #10	CRN 213*	#161	#159	CRN 136†	KC 98-16	PHN 4085‡	PHN 4034‡
SiO_2	51.09	53.29	51.72	54.16	52.02	46.53	46.83	41.03	40.56	40.20	43.01	43.98	44.06
TiO_2	0.60	0.27	0.72	0.33	0.29	0.36	0.08	3.20	4.09	3.04	0.91	0.87	2.02
Al_2O_3	6.07	3.88	5.37	3.50	4.23	12.52	12.59	16.00	15.60	14.71	13.18	14.29	13.44
Cr_2O_3	0.01	0.00	0.00	0.04	1.07	1.57	1.04	0.01	0.05	0.00	1.58	1.67	1.56
FeO	12.97	13.55	12.86	9.43	2.60	4.67	5.52	10.01	10.87	10.40	4.73	4.38	4.50
MnO	0.42	0.46	0.41	0.17	0.07	0.38	0.30	0.16	0.16	0.16	0.06	0.06	0.00
MgO	26.52	27.29	26.80	29.51	15.68	14.14	13.84	14.54	14.27	14.62	18.18	17.80	17.33
CaO	1.86	1.92	1.80	1.38	20.3	18.06	18.11	9.75	9.45	9.62	10.16	10.14	10.12
Na_2O	0.05	0.13	0.00	0.29	2.23	1.09	0.47	3.04	3.29	3.16	4.53	4.10	3.40
K_2O	0.00	0.00	0.00	0.00	0.11	0.04	0.02	1.64	1.14	1.16	0.63	0.51	1.41
Total	99.59	100.79	99.68	98.81	98.60	99.36	99.04	99.38	99.48	97.07	97.85	97.80	97.84
$Mg/(Mg+Fe^{2+})$	0.785	0.782	0.788	0.848	0.915	0.844	0.817	0.721	0.701	0.715	0.915	0.879	0.873
$Ca/(Ca+Mg)$	0.048	0.048	0.046	0.033	0.482	0.479	0.485	0.325	0.322	0.321	0.478	0.290	0.296

*Aluminous clinopyroxene of metasomatic origin.   †Inclusion in a garnet megacryst.   ‡Metasomatic amphibole.

mineral phases from the alnöites demonstrated that the maximum depth of equilibration recorded was ~120 km (~3.6 GPa), with a maximum temperature of ~1250°C (*14–18*). Hence, the alnöite pipes were described as barren with regard to diamond potential. Using empirical geobarometers, Collerson *et al*. (*1*) reported pressures between 5 and 22 GPa (165 and 730 km). When pressures for mineral phases from the garnet-spinel peridotites (*8–9*) are calculated using these empirical geobarometers, assuming that the clinopyroxene is majorite, they give pressures of between 5 to 23 GPa (165 and 760 km). Garnet gives the lower pressure (5 to 7 GPa, or 165 to 230 km), and clinopyroxene, now interpreted as majorite, gives the higher pressures (18 to 23 GPa, or 590 to 760 km)—even if they are from the same sample. A similar bimodality of pressures is seen between garnet (6.2 GPa, or 205 km) and "majorite" (13.6 to 20.9 GPa, or 450 to 690 km) compositions from the same sample, KC-98-16, reported in table 1 of (*1*). Secondary clinopyroxene and metasomatic amphibole in the garnet-spinel xenoliths (*8–12*; Fig. 2, A and B), if also interpreted as majorite, yield pressures of 7 to 23 GPa (230 to 760 km).

The previously reported data could conceivably be interpreted in terms of the Collerson *et al*. (*1*) model of a deep mantle origin. However, these data are from xenoliths that contain both garnet and spinel, with the former rimming the latter (Fig. 2A), and that therefore must have been derived close to the garnet-spinel transition in the upper mantle. It has previously been demonstrated [e.g., (*36, 37*)] that this transition is gradational, being dependent upon the mineralogy and trivalent ion content (especially Cr_2O_3), and a pressure range for the coexistence of garnet

and spinel between 1.2 and 3.1 GPa (40 and 100 km) has been reported [e.g. (*38–40*)]. Additionally, the secondary clinopyroxene and amphibole (Fig. 2, A and B) in the garnet-spinel xenoliths is retrograde in nature (*8–11*). Yet, within the area displayed in the figure, pressures calculated using the geobarometers of (*1*)—assuming a garnet stoichiometry for all phases—range from 5 to 21 GPa (165 to 690 km).

Textures. The textures reported by (*1*) showed a "majorite"-spinel symplectite rimming a zone of kelyphite that in turn rimmed a pyrope-rich garnet [sample KC-98-16; figure 3b of (*1*)]. If majorite is indeed present, this very-high-pressure phase rims low pressure kelyphite, which in turn rims moderate pressure pyrope. Collerson *et al*. interpreted these relationships as showing that the "majorite"-spinel symplectite predated the glassy kelyphite rim of the garnet (*1*). If so, however, why did the majorite remain unaffected while the garnet broke down during ascent in the alnöite to produce kelyphite? Sample KC-98-16 appears actually to be a typical lherzolite and a classic example of the transformation from the garnet to spinel peridotite facies according to the reaction $Mg_3Al_2Si_3O_{12}$ + $Mg_2SiO_4 = 2Mg_2Si_2O_6 + MgAl_2O_4$ (*36, 37*). The interpretation of majorite in the texturally secondary spinel assemblage, therefore, is not a unique explanation for the mineral chemistry. In our view, the reaction above better explains the textures displayed by this sample [figure 3, B, C, D, and E, of (*1*)], in that the assemblages depicted would represent a consistent pressure decrease (or temperature increase) from the central pyrope to what would be a pyroxene-spinel symplectite. Similar textures occur in symplectitic

and cellular garnet breakdown assemblages (pyroxene + spinel) in crustal granulites (*35*) and in eclogites (*34*). Indeed, peridotite xenoliths from Malaita exhibit garnet breakdown textures manifest as clinopyroxene-spinel symplectites (Fig. 2C). Furthermore, if the discrete primary pyroxenes are perovskite-structured, then monoclinic diopside should exhibit the parallel extinction of orthorhombic Ca-Mg perovskite (*21–24*). Coexisting olivine in KC98-16 should be ringwoodite (cubic) and optically isotropic. Neither x-ray data nor optical observations are given, a deficiency that dramatically weakens the interpretation of the mineral assemblages as originating in the transition zone.

Majorite is preserved only in transiently shocked meteorites, rapidly quenched experiments from high *P-T* conditions, and pyroxene solid solution of garnet trapped in some diamonds and rare xenoliths (*21–24*). Majoritic garnet in xenoliths and in some diamonds (*26–27*) decomposes to garnet + crystallographically controlled pyroxene along cubic garnet planes. Kinetics are the key to preservation—and even in diamond and under the relatively rapid intrusion rates postulated for explosive mantle melts (*41*), retrograde reactions occur. The anastomosing veins reported to contain majorite in the Malaita garnets are open-system highways that are unlikely to preserve majorite or silicate perovskite. Indeed, the E type majorite that occurs in veins or randomly distributed amoeboid blebs have compositions similar to metasomatic amphibole in the garnet-spinel and spinel peridotite xenoliths [(*19*) and Table 2]; the relatively Si-rich majorites are reported to have low first-order birefringence, again

Table 3. Comparison of "perovskite" mineral compositions (*1*) with similar compositions in the upper mantle suite (*14–20*) from Malaitan, Solomon Islands.

Mineral:	"Mg-perovskite"			"Ca-perosvkite"						
Rock (ref.):	P type (*1*)	Peridotite (*8–13*)		E type (*1*)		Mega. (*8–13*)	P type (*1*)		Peridotite (*8–13*)	
Sample #:	KC 98-16	CRN 209*	CRN 215*	KC 98-8	KC 98-17A	CRN 60†	KC 98-16 #23	KC 98-15 #29	PHN 4013‡	PHN 4069‡
SiO_2	53.26	54.95	55.13	44.37	48.04	49.27	51.50	52.61	50.52	51.47
TiO_2	0.05	0.24	0.08	2.48	2.58	0.50	0.29	0.30	0.26	0.30
Al_2O_3	4.57	4.30	4.25	12.44	6.07	8.78	4.09	4.15	6.40	6.00
Cr_2O_3	0.45	1.10	0.34	0.05	0.07	0.09	1.05	1.09	1.05	0.81
FeO	7.67	5.55	6.74	7.54	7.14	7.23	2.61	2.59	4.51	4.16
MnO	0.42	0.10	0.14	0.18	0.28	0.34	0.08	0.07	0.19	0.30
MgO	30.10	31.64	31.80	11.48	14.75	15.88	15.77	15.53	15.77	16.11
CaO	1.58	1.74	0.48	19.05	20.63	16.83	20.13	20.32	19.89	19.70
Na_2O	0.20	0.00	0.00	1.99	0.26	1.12	2.33	2.28	1.22	1.14
K_2O	0.07	0.00	0.00	0.17	0.02	0.00	0.00	0.00	0.00	0.00
Total	98.37	99.62	98.96	99.75	99.84	100.04	97.85	98.94	99.81	99.99
Mg/(Mg+Fe²⁺)	0.875	0.910	0.894	0.731	0.786	0.796	0.915	0.914	0.862	0.873
Ca/(Ca+Mg)	0.036	0.038	0.011	0.544	0.501	0.432	0.478	0.485	0.475	0.468

*Primary orthopyroxene in peridotite xenoliths.   †Inclusion in a garnet megacryst.   ‡Secondary clinopyroxene "sliver" between garnet and spinel in the garnet-spinel peridotites.

similar to the metasomatic amphibole (*18, 19*). If the mineral is majorite, all should be optically isotropic. While symmetry changes in majorite from cubic to tetragonal, for Si-rich majorite (between 3.8 and 4 Si atoms per 12 oxygens), have been observed, they are exclusively from high-pressure experiments in the simple synthetic Mg-Al-Si system [(*42*) and references therein]. In the complex Na-Ca-Fe-Mg-Cr-Al-Si system, which is relevant here, symmetry changes of this type have not been described.

Infrared spectroscopy. In an attempt to identify majorite and Mg-perovskite, Collerson *et al.* (*1*) used infrared spectroscopy in reflectance mode, which can be ambiguous, especially if unpolarized. Before any interpretation, one has to understand orientation and surface effects on the spectra produced. It is not clear that
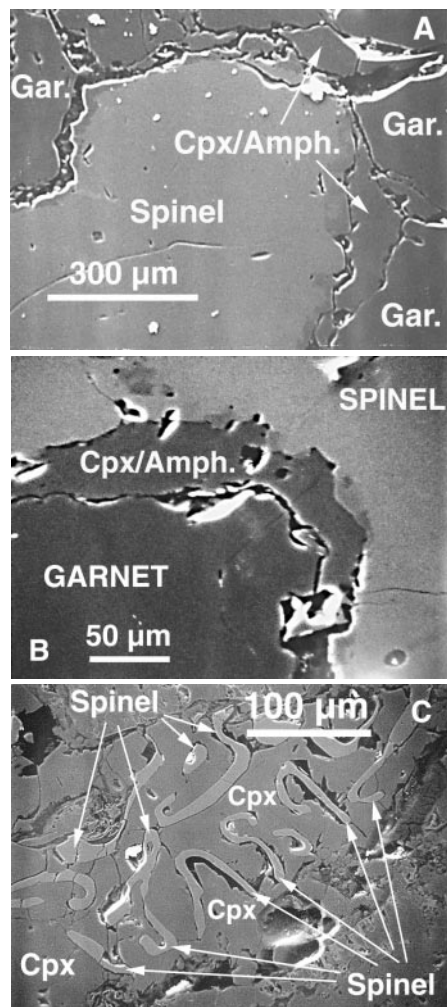

Fig. 2. Photomicrographs of the mineral textures in the garnet-spinel peridotite xenoliths (*14–20*). (**A**) Back-scattered electron (BSE) image of garnet and spinel relationships; retrograde amphibole or clinopyroxene is observed between the garnet and spinel. (**B**) BSE image showing a close-up of the relationships described in Fig. 2A. (**C**) BSE image showing a spinel-clinopyroxene symplectite that resulted from the breakdown of garnet.

this has been done. Collerson *et al.* attributed the broadening of the peak in Mg-perovskite, relative to the standard pyroxenes, to disorder, and that interpretation is not unrealistic. Disordered material with pyroxene stoichiometry does not necessarily have "Mg-perovskite chemistry," however. Any kind of complex solid solution (Ca-Fe-Mg) may induce disorder. Therefore, peak broadening reflects the chemical complexity of the analyzed material rather than providing information on its structure.

Moreover, the peaks at 1050 cm^{-1} and 750 cm^{-1}, typical of Si-O-Si bridges, are consistent with Si in the tetrahedral position. Therefore, the analyzed mineral cannot be a perovskite-structured mineral, because no Si in the octahedral position has been detected. Why were the spectra of synthetic silicate perovskites [e.g., (*43*)] not used for comparison? We note that cation/anion coordination is measurable with the Raman microprobe; such measurements should be done both for reported "majorite" and "perovskite." Indeed, we could argue that the spectra of "the phase with majorite chemistry" in figure 5B of (*1*) constitute a much closer fit to the pyroxenes in figure 5a of (*1*)—which is not surprising, because Mg-perovskite has a pyroxene chemistry (Table 3).

In summary, the interpretation of the mineral data reported by Collerson *et al.* (*1*) as representing the transition zone and the uppermost lower mantle is premature. The presence of all but two of the reported mineral compositions in xenoliths or xenocrysts of proven upper mantle provenance demonstrates that the interpretation in (*1*) is non-unique. Without definitive x-ray data, we would interpret the data reported by Collerson *et al.* (*1*) as being from the upper mantle and consistent with previous reports of the mantle xenoliths from the Malaitan alnöites (*14-20*). The deep-mantle interpretation for the mineral compositions reported by Collerson *et al.* (*1*) remains highly equivocal.

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41. Ascent times can be weeks to months in duration (which is more than sufficient for transformation) through the upper mantle and lower crust, but only seconds to minutes close to the surface (*44*).
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Response: We welcome the opportunity to elaborate on our previous report of transition zone macrocrysts and xenoliths from Malaita, particularly with regard to speculative mantle models for the Ontong Java Plateau. Before we reply to the comment, however, we wish to reject the assertion, implicit in the comment by Neal *et al.*, that we did not sufficiently discuss the ambiguity of identifying mineral phases based on chemistry alone in our original study. Indeed, we concluded [page 1217 of (*1*)] that Mg-tschermakitic orthopyroxene could be misidentified as low-Ca majoritic garnet, which makes up ∼99% of the studied majorite chemistries [pages 1216 and 1217 of (*1*)]. The only debate, therefore, centers on whether our reported low-Ca majoritic chemistries are in fact orthopyroxenes and whether the mantle models cited by Neal *et al.* provide new evidence in favor of a low-pressure origin.

Neal *et al.* remain unconvinced of the majoritic and perovskitic nature of our silicate analyses, and call for "definitive x-ray data" to seal the argument. In doing so, they expect us to do the physically impossible. A number of single-crystal x-ray diffraction studies on diamond-hosted silicate inclusions of ultrahigh pressure chemistry have failed to detect the original crystallography but invariably have found that even in the inert diamond, isochemical phase transformation has led to reversion of the high-pressure phases due to plastic deformation during kimberlite eruption (*2–4*). These problems are amplified for ultradeep minerals that are not contained in diamond, particularly in view of fast decompression. We dealt with this phenomenon in (*1*); our spectroscopic characterizations of these materials demonstrated that they were disordered silicates, as expected for high-pressure phases that had reverted on decompression. Significantly, ultrafast emplacement of the Malaita "alnöite" has since been confirmed by Ar-Ar chronology (*5*). In brief, there is no a priori reason to expect that the Malaita majorites and perovskites should have preserved their original crystallography; indeed, the opposite is expected (*1–4*). Mineral chemistry remains our main source of information. That tool is widely accepted for identification of diamond-hosted silicate inclusions (*2–4*) and, we maintain, is also valid for macrocryst and xenolith studies.

Neal *et al.* propose that there is a strong compositional similarity between previously reported mineral chemistries from Malaita xenoliths and those that we described as majorites, a point supposedly illustrated in their figure 1. (It is worth noting that, to our knowledge, there exist no "definitive x-ray data" that would confirm the identification by Neal *et al.* of those minerals as being pyroxenes and amphiboles.) As we noted in (*1*), however, the vast majority of our samples have low Ca-contents and could therefore not be confused with clinopyroxene

and the type of amphibole reported in table 2 of Neal *et al.* Further, an inspection of the orthopyroxene data in their figure 1 reveals not only that the supposed colinearity on that diagram with the majorite chemistries that we reported is far from excellent, but also that the most aluminous orthopyroxenes could not nearly explain the full range in chemistries. Orthopyroxenes from garnet-peridotite and bronzite megacrysts plot at the most siliceous end of the spectrum—≥3.8 Si per garnet formula unit (pfu). Although not evident from their figure 1, the compositional range between 3.8 and 4.0 Si pfu was found to be devoid of majorites by us. This is best illustrated in figure 1A of (*1*). It is also noteworthy that, contrary to the assertion of Neal *et al.*, the bronzite data plot to the left of the majorite vector, as do other orthopyroxene compositions that we had plotted for comparison in our original study [figure 1C of (*1*)]. Thus, the only orthopyroxene data of relevance are the data points compiled by Neal *et al.* from spinel peridotite (open crosses in their figure 1). These data range from 3.95 to 3.75 Si pfu, a marginal overlap with the most siliceous data points reported in (*1*). We never argued that the most siliceous data points could not, in theory, be orthopyroxenes. Nonetheless, we observe that data points with Si greater than 3.85 were conspicuously lacking in our data set (*1*), a fact for which Neal *et al.* provide no explanation.

More important, our original study (*1*) commented on this possible ambiguity to explain the less siliceous compositions as orthopyroxenes. Specifically, we cited experimental work that suggested that highly aluminous orthopyroxene required low pressure and high temperature (>1500°C) for equilibration [reference 20 in (*1*)]. Judging from the available orthopyroxene data that we compiled, we find our case confirmed in that there is a patent lack of compositions with Si pfu <3.75. That absence, in turn, suggests normal upper mantle temperatures that would not have allowed the formation of highly aluminous orthopyroxene that could be confused with majorite. Furthermore, geothermometry of garnet-peridotite, garnet-spinel peridotite, and macrocrysts from Malaita has previously indicated a maximum equilibration temperature of ∼1250°C, as summarized by Neal *et al.*; such normal temperatures again appear to be supported by the lack of highly aluminous orthopyroxene in these peridotites.

We are puzzled by the evident change of mind shown by Neal *et al.*, who now state that the plume head responsible for the generation of the Ontong Java Plateau may have been 400°C hotter than ambient mantle, thereby allowing development of pyroxenes with a large tschermak component. We prefer to base our interpretation on actual observations and conclude that at the observed maximum tempera-

ture, orthopyroxene could not become sufficiently aluminous (as appears to be supported by observed mineral chemistry) to be confused with majorite. To summarize, the operative point is that our majoritic compositions cannot be explained by the presence of compositionally variable othropyroxene in our samples. The compositional trends and ranges for majorite reported by Collerson *et al.* (*1*) are correct, and the thermal conditions are not supported by available geothermometry summarized by Neal *et al.* The principal commonality between the two materials is simply that both majorite and orthopyroxene are each magnesium silicates with variable (but diagnostic) chemistries. It is that simple commonality that underlies the confusion shown by Neal *et al.*

Finally, Neal *et al.* use their comment to publicize a mantle model for the Ontong Java Plateau. If we have interpreted this model correctly, they claim that two pulses of Ontong Java magmatism (at ∼122 and ∼90 Ma) were associated with the surfacing of a plume head, which left a compositional root beneath the plateau. Macrocrysts that were later brought to the surface by alnöite are seen as fractional-crystallization products of the original magma pulses. Neal *et al.* suggest that our samples formed or equilibrated in the deeper (>100 km) portion of the plume head, but not as deep as the transition zone. They cite work by Richardson *et al.* (*6*) and material presented as abstracts or in the press as supporting evidence for their model, in that the mantle beneath the plateau might be compositionally different from asthenosphere to depths of 150 to 200 km.

We agree that the creation of the Ontong Java Plateau required a plume and that the flattening of the original plume head beneath the lithosphere would have modified the lower part of the preexisting oceanic lithosphere and asthenosphere. However, we fail to recognize the importance of that scenario for the interpretation of the Malaita macrocrysts. Clearly, the alkaline magmatism at ∼35 Ma, to which the Malaita alnöites belong, is unrelated to the Ontong Java plume and reflects passage of the Ontong Java Plateau over a separate mantle hotspot (*7*). Alnöite emplaced during this event entrained majorite-bearing assemblages from the mantle transition zone as well as shallower peridotite lithospheric xenoliths. In this respect, Collerson *et al.* (*1*) cited seismic tomographic evidence (*8*) for the southwestern Pacific region that shows an extended zone of high seismic velocities in the transition zone, probably formed by accumulation of subducted oceanic crust. Currently available seismic information is unlikely to provide compelling evidence for either of these competing models. Determination of Pb-isotope ratios of the garnetites might provide an answer, because the U-Th-Pb system appears to have the best chance (*9*) of distinguishing be-

tween relatively young fractional-crystallization products (the model of Neal *et al.*) and potentially ancient subducted material (our model). In this context, it is crucial to note that in the one example for which Pb-isotope data are now available (*9*), megacryst formation by fractional crystallization from proto-kimberlite had to be abandoned as a potential origin for such samples.

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