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A negative Ce anomaly in a peridotite xenolith: Evidence for crustal recycling into the mantle or mantle metasomatism?

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Abstract—The presence of negative cerium anomalies in island arc lavas has been considered enigmatic. Such negative Ce anomalies must be inherent in the source region and can be produced by subducting pelagic sediments + seawater-altered basalts (SWAB) into the mantle. A mantle peridotite from the Malaitan alnöite also contains a negative Ce anomaly, which can be produced by sediment recycling into the upper mantle. However, in spite of the poorly defined effects of cryptic metasomatism and associated fO_2 conditions, such a process also seems plausible for the generation of negative Ce anomalies.

In order to define the petrogenesis of this peridotite within the constraints of the present experimental data, we have attempted various mixing models with the end-members: mantle peridotite, Pacific sediment and seawater-altered basalt. In our model, it was assumed that negative Ce anomalies cannot be produced by magmatic or metasomatic processes. Best results were obtained from mixing a maximum of 3 to 5% seawater-altered basalt and 2 to 3% pelagic sediment, with a LREE-enriched mantle precursor. Our model stresses the importance of the contribution that recycled crustal materials can have on the composition of the upper mantle, in particular the recognition of a crustal signature in a mantle regime.

INTRODUCTION

THE RARE EARTH ELEMENTS (REE) have proven to be effective geochemical indicators, since they are sensitive to igneous processes that operate during magma genesis and evolution (*e.g.* ALLÈGRE *et al.*, 1977; ALLÈGRE and MINSTER, 1978; HANSON, 1980). The configuration of the 4f electron shell allows Ce and Eu to deviate from the general 3+ valency. These elements may develop anomalies due to reduction ($Eu^{3+} \rightarrow Eu^{2+}$) or oxidation ($Ce^{3+} \rightarrow Ce^{4+}$). It is the vastly different dissociation constants (Kd) of the reduced and oxidized species that produce anomalies in otherwise smooth REE profiles. In igneous systems, the Eu^{2+} species may already exist at the onset of partial melting, but it is possible that reduction of Eu^{3+} can occur in magmatic processes. The increased ionic radius of Eu^{2+} and its compatible charge leads to its preferential incorporation (Kd Eu^{2+} > Kd Eu^{3+}) into the Ca site of plagioclase (Kd_{plag} Eu^{2+} > Kd_{plag} Eu^{3+}). As such, negative and positive Eu anomalies have been reported in many igneous rocks including meteorites, lunar samples, as well as terrestrial examples (*e.g.* WARREN, 1985; HENDERSON, 1984).

The presence of Ce anomalies in igneous rocks is enigmatic. The experimental work of SCHREIBER *et al.* (1980) demonstrated that Ce^{4+} is unlikely to occur in a magmatic system due to the reaction

$$Ce^{4+} + Fe^{2+} = Ce^{3+} + Fe^{3+}.$$

As Fe^{2+} will always be in excess of Ce in basaltic magmas, the presence of Ce^{4+} should never occur in magmatic situations. With Ce present in only one oxidation state in the magma (*i.e.* only one Kd for Ce ions), no igneous process is available for the production of an anomaly.

Negative Ce anomalies in basaltic island arc lavas have been reported from a variety of locations [*e.g.* the Solomon Islands (JAKES and GILL, 1970; RAMSAY *et al.*, 1984), Papua New Guinea (HEMING and RANKIN, 1979) and the Mariana Islands (DIXON and BATIZA, 1979; WHITE and PATCHETT, 1984)]. HEMING and RANKIN (1979) and HOLE *et al.* (1984) argued for the negative Ce anomaly to be present in the source region prior to partial melting. However, another hypothesis is that Ce anomalies arise in the mantle through fluid-rock fractionation associated with the slab (WHITE, pers. commun., 1989). Evidence for this comes from areas where arc lavas contain slight negative Ce anomalies, whereas the sediment immediately in front of the arc contains a bulk positive Ce anomaly (WHITE and DUPRE, 1986).

ORIGIN OF NEGATIVE Ce ANOMALIES

The behavior of Ce is largely controlled by its oxidation and reduction chemistry. The oxidation of the soluble Ce^{3+} to the insoluble Ce^{4+} species is responsible for the negative Ce anomaly in seawater (*e.g.* ELDERFIELD and GREAVES, 1982; PALMER, 1983; DEBAAR *et al.*, 1983, 1985). Scavenging of elements by settling Fe-Mn particles facilitates the preferential removal of Ce^{4+} from the water column and incorporation into Fe-Mn nodules which contain large positive Ce anomalies (ELDERFIELD *et al.*, 1981; FLEET, 1984). The REE contents of pelagic sediments indicate that there is usually a negative Ce anomaly present (*e.g.* PIPER and GRAEF, 1974; HOLE *et al.*, 1984; WANG *et al.*, 1986), due to interaction with seawater. Furthermore, the magnitude of the Ce anomaly in pelagic sediments appears to vary with age (WANG *et al.*, 1986; KAY and KAY, 1988). A negative Ce anomaly has also been observed in seawater-altered basalts (SWAB) (*e.g.* MASUDA and NAGASAWA, 1975; MENZIES *et al.*, 1977; LUDDEN and THOMPSON, 1979), again due to interaction with seawater. Therefore, it is evident that certain areas of the seafloor can exhibit bulk negative Ce anomalies.

Several studies have reported negative Ce anomalies in island arc lavas, but did not comment upon them or considered them a weathering phenomenon or an analytical artifact (*e.g.* MASUDA, 1968; TAYLOR *et al.*, 1968; JAKES and GILL, 1970; Ewart *et al.*, 1973; Kay and Hubbard, 1978). Later authors proposed that the introduction of pelagic sediment into the source region can produce such anomalies (Heming and Rankin, 1979; Dixon and Batiza, 1979; White and Patchett, 1984; Hole *et al.*, 1984).

Therefore, it is significant to discover a signature of an inherent negative Ce anomaly in a mantle xenolith from the Malaitan alnöite, Solomon Islands. This implies a significant role for crustal recycling in the evolution of the mantle and generation of magmas. Indeed, Bielski-Zyskind *et al.* (1984), in an isotopic study of Malaitan mantle xenoliths, concluded that a crustal component existed in the mantle beneath Malaita. Frey and Green (1974) have also reported negative Ce anomalies from three lherzolite xenoliths entrained in basanite from Victoria, Australia. Even the host basanite has a negative Ce anomaly. However, these authors neglected to comment upon these REE patterns, possibly because of the errors inherent in analysis by INA. Ottonello *et al.* (1979) reported negative Ce anomalies in some Ligurian peridotites from the Alps. These authors considered the Ce anomaly to be a product of peridotite serpentinization, but report only whole-rock data, rather than clinopyroxene separates.

We report the occurrence of a negative Ce anomaly in a mantle peridotite xenolith (CRN211) from the Malaitan alnöite, Solomon Islands. The REE contents of this peridotite form the main thrust of this paper. Major and trace element, and isotope geochemistry of the minerals of this peridotite confirm its mantle origin (Neal, 1985). The presence of this Ce anomaly has important implications for crustal recycling into the mantle beneath Malaita and generation of Ce-anomalous lavas (*cf.* Jakes and Gill, 1970; White and Patchett, 1984).

GEOLOGICAL SETTING

The Solomon Islands chain delineates the boundary between the Pacific and Indo-Australian plates. The area is dominated by the Ontong Java Plateau (OJP), which is an overthickened portion (up to 42 km) of oceanic crust (Coleman, 1976; Hussong *et al.*, 1979), abutting the Indo-Australian plate (Fig. 1). The island of Malaita is formed from the obducted leading edge of the OJP. Pipe-like bodies of alnöite (a mica lamprophyre with melilite in the groundmass) were explosively emplaced in limestones and mudstones which have been folded into NW–SE trending anticlines and synclines (Rickwood, 1957). The pipes at Babaru'u and Kwaikwai have a core of fine-grained, black alnöite containing megacrysts, surrounded by an autolithic breccia containing peridotite xenoliths, macrocrysts, megacrysts and xenoliths of crustal country rock (Nixon and Boyd, 1979; Neal, 1985).

RARE EARTH ELEMENTS

Analytical technique

An ultrapure mineral separate of clinopyroxene from CRN211 was prepared by progressive leaching in HCl and HF acids with repeated handpicking. Whole-rock analysis was considered meaningless due to the degree of olivine alteration. The ultrapure clinopyroxene mineral separate was analyzed by isotope dilution mass spectrometry for the REE and was spiked prior to dissolution. The REE were separated on ion exchange columns and measured automatically on a VG Isomass 54E spectrometer, using the method of Thirlwall (1982). The terms "depleted" and "enriched" are used relative to a flat, chondrite-normalized, bulk Earth REE pattern.

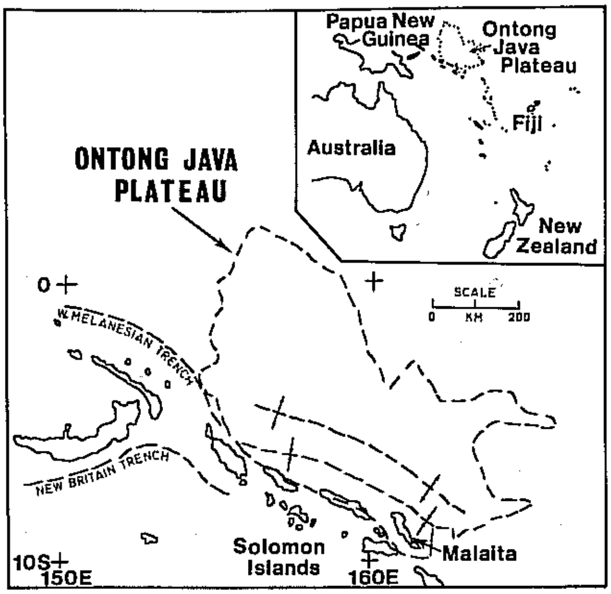

FIG. 1. Tectonic sketch map of the SW pacific region.

The analytical technique used is critical in the detection of negative Ce anomalies. Such anomalies could result from the dissolution procedure, if Ce was oxidized and precipitated prior to equilibration with the spike (Thirlwall and Graham, 1984). We feel this is unlikely as the sample was spiked prior to dissolution and care was taken to ensure that all sample was dissolved.

Results

The REE analysis of CRN211(cpx) is presented in Table 1 and Fig. 2, and a large, negative Ce anomaly is superimposed on a generally LREE-enriched profile ($[Ce/Ce^*]_N = 0.36$). The remaining clinopyroxene separate was releached and handpicked prior to two further analyses for the REE. Results are within ±1% of the first REE profile. Based on the modal % of clinopyroxene in the peridotite and assuming all the REEs are contained in clinopyroxene, the whole-rock pattern may be calculated (Fig. 2). REE contents of the olivine, orthopyroxene and spinel are typically so low as to not raise these whole-rock estimates appreciably (*i.e.* <5%).

MODELING

Although there is compelling evidence to suggest the subduction of Ce-anomalous crustal components, there is a lack of experimental evidence to either support or refute the possibility of a metasomatic or fluid-rock interaction for the production of negative Ce anomalies. In our study, we model

TABLE 1: Rare earth element abundances in the clinopyroxene from CRN211.

	$CRN211_{(cpx)}$
La	7.25
Ce	6.17
Nd	10.2
Sm	2.56
Eu	0.89
Gd	2.72
Tb	(0.51)
Dy	2.48
Er	1.14
Yb	0.83
Lu	0.11

Brackets indicate estimated abundance.

the xenolith CRN211 by assuming sediment subduction into the mantle. This is not to suggest this is the only mechanism by which such negative Ce anomalies can be produced, but at present it is the most constrained. Furthermore, negative Ce anomalies in subducted pelagic sediments and SWAB could *locally* impart a negative Ce anomaly to the mantle, if it survived the subduction processes.

In order to explain the negative Ce anomaly in mantle peridotite CRN211, we have attempted bulk mixing calculations between mantle and subducted components. Similar modeling has been undertaken by HOLE *et al.* (1984) to generate the Ce-anomalous lavas from the Mariana Islands. These authors used a MORB-depleted mantle, Pacific Authigenic Weighted Mean Sediment (PAWMS—an average of Cenozoic sediments) and a small fluid contribution (1%) from the dehydrating slab. The PAWMS component is comprised of 95% average nanofossil ooze and 5% average ferruginous clay from the Nazca plate. They considered that a contribution between 0.3 and 0.5% PAWMS was required in the source to generate the small negative Ce anomalies. The small fluid contribution is required to generate high Rb/Ba ratios.

We have not considered this fluid contribution, as we assume that the generation of the mantle peridotite occurs at a deeper level, at which the slab-derived fluid will be of minimal importance. Our modeling includes: 1) a seawater-altered basalt (SWAB) component; 2) pelagic sediment; and 3) mantle peridotite. The SWAB component is included because this forms a significant proportion of subducted material and can contain a negative Ce anomaly of its own (*cf.* LUDDEN and THOMPSON, 1979).

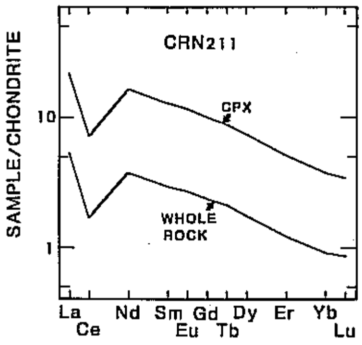

FIG. 2. Chondrite normalized rare earth element patterns of $CRN211_{(CPX)}$ and $CRN211_{(WR)}$. The whole-rock pattern has been calculated from the modal analysis of CRN211 assuming all REE are contained in cpx. Modes: 54% olivine; 24% clinopyroxene; 14% orthopyroxene; 5% spinel; 3% kelyphite. Normalizing values are from NAKAMURA (1974).

TABLE 2: Modelling components presented in Figure 4.

	PHN4015	PAWMS	A-29	SWAB
La	0.41	25.8	21.3	15.8
Ce	1.39	9.60	13.6	11.9
Nd	1.40	19.3	28.9	14.6
Sm	0.43	4.40	4.84	4.13
Eu	0.15	1.13	1.06	1.33
Gd	0.43	7.72	(3.73)	(4.22)
Tb	(0.07)	0.83	0.78	0.74
Yb	0.07	5.55	2.39	2.87
Lu	0.01	0.62	0.40	0.49

Brackets indicate estimated abundances. Peridotite PHN4015 from Neal (1985); PAWMS composition of Hole et al. (1984); A-29 = Pacific sediment from Piper and Graef (1974); SeaWater-Altered Basalt (SWAB) composition 15-2A of Ludden and Thompson (1979).

End-member components

We assume that the mantle component contains no Ce anomaly and must be LREE-enriched in order to produce the general LREE enrichment of our mantle peridotite. The LREE-enriched Malaitan peridotite PHN4015 is taken as our mantle component (Table 2 and Fig. 3) for the modeling of CRN211. This is representative of the LREE-enriched mantle portion that is present beneath Malaita (NEAL, 1985, 1988). Therefore, the sediment-SWAB component must contain a bulk negative Ce anomaly greater than that observed in CRN211.

The modeling components must also predate the age of alnöite eruption (34 Ma; DAVIS, 1977), being at least early Eocene in age. Two sediment compositions, displaying large, negative Ce anomalies, are used (Table 2 and Fig. 3): 1) the

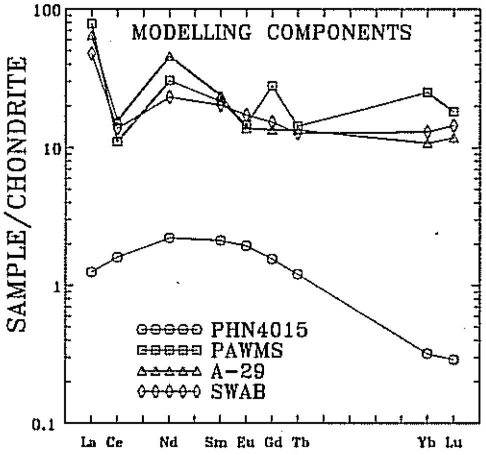

FIG. 3. Components used in the modeling of peridotite CRN211 represented on a chondrite-normalized REE plot. Normalizing values are from NAKAMURA (1974). Data sources are: PHN4015 = NEAL (1985); Pacific authigenic weighted mean sediment (PAWMS) = HOLE *et al.* (1984); Pacific sediment A-29 = PIPER and GRAEF (1974); Seawater-altered basalt (SWAB) composition 15-2A of LUDDEN and THOMPSON (1979).

Cenozoic PAWMS composition of HOLE *et al.* (1984) ($[Ce/Ce^*]_N = 0.21$); and 2) the surface sediment A-29 from the East Pacific Rise of PIPER and GRAEF (1974) ($[Ce/Ce^*]_N = 0.27$). These components are chosen in order to demonstrate that the REE profile of CRN211 may be generated by subduction and mixing of SWAB and pelagic sediment with a LREE-enriched mantle.

There is some controversy as to the presence and magnitude of a negative Ce anomaly in pre-Cenozoic Pacific sediments. KAY and KAY (1988) indicate the pre-Cenozoic northern Pacific pelagic sediments contain only small negative Ce anomalies ($[Ce/Ce^*]_N \cong 0.8$). However, LIU *et al.* (1988) described upper Cretaceous carbonates with large negative Ce anomalies from DSDP Hole 316 ($[Ce/Ce^*]_N \cong 0.2$). In this respect, the PAWMS composition of HOLE *et al.* (1984) is a feasible component involved in the petrogenesis of our mantle peridotite. The SWAB component 15-2A (Table 2 and Fig. 3) of LUDDEN and THOMPSON (1979) is also used in our modeling. This sample is the weathered outer portion of an Atlantic MOR basalt, and although it is a somewhat extreme composition, it is of the required age (46 Ma) and contains the requisite negative Ce anomaly ($[Ce/Ce^*]_N = 0.38$). It is the unusual REE signature of CRN211 which suggests that extraordinary and somewhat unique conditions are required for its petrogenesis. Therefore, the crustal components chosen are, by necessity, somewhat extreme.

TABLE 3: Modelling results.

	A 94% Mantle 3% SWAB 3% PAWMS	B 93% Mantle 5% SWAB 2% A-29
La	1.63	1.60
Ce	1.95	2.16
Nd	2.33	2.61
Sm	0.66	0.70
Eu	0.21	0.23
Gd	0.76	0.69
Tb	0.11	0.12
Yb	0.32	0.26
Lu	0.04	0.04

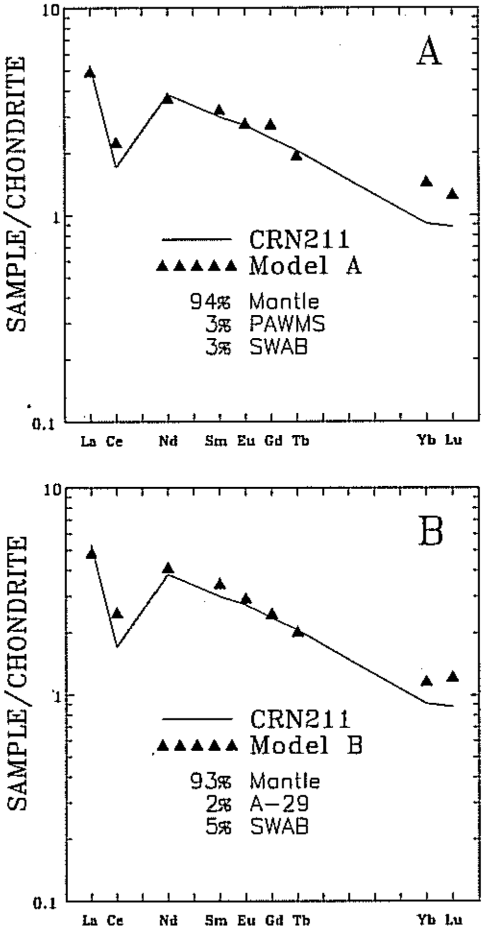

FIG. 4. Results of the modeling undertaken in order to generate CRN211 represented on a chondrite normalized REE plot. Normalizing values are from NAKAMURA (1974). The filled triangles are the modeling results, and solid lines represent the components used. Proportions required for the "best fit" in each case are also given. Note that this figure is meant to be illustrative rather than absolute.

Modeling results

The modeling presented in Fig. 4 and Table 3 is intended to be illustrative rather than absolute. These results demonstrate that our mantle peridotite can be generated by mixing various combinations of subducted components with a LREE-enriched mantle: a) sediment + PHN4015; b) SWAB + PHN4015; or c) both sediment and SWAB + PHN4015. It is difficult to estimate which combination of components is the most realistic, inasmuch as all sediment may be removed by obduction, and the SWAB component may not always exhibit a negative Ce anomaly (*e.g.* LUDDEN and THOMPSON, 1979). However, the amount of pelagic sediment required is greater than the <1% postulated for the petrogenesis of island arc volcanics (*e.g.* HOLE *et al.*, 1984; WHITE and PATCHETT, 1984; KAY and KAY, 1988).

DISCUSSION

The pronounced negative Ce anomaly exhibited by mantle peridotite CRN211 suggests pelagic sediment/SWAB input by subduction into the mantle. Such a process has been proposed in order to account for the compositions of some ocean island basalts (*e.g.* HOFMANN and WHITE, 1982). It is extremely likely that a proportion of pelagic sediment will be subducted, but is it all incorporated into the island arc volcanics? The results of this study suggest not.

It is not clear whether metasomatism could have produced the negative Ce anomaly. Metasomatic fluids are hot, supercritical fluids containing large quantities of Fe during *patent metasomatism* (*e.g.* DAWSON, 1984; MENZIES and HAWKESWORTH, 1987) which would inhibit the formation of the insoluble Ce^{4+} species, negating the formation of a negative Ce anomaly by this process. However, the nature of *cryptic metasomatic* fluids is more vague. The general LREE-enriched signature of CRN211 could have been imparted by cryptic metasomatism, but whether such a process produced the negative Ce anomaly is unclear. Although the experiments of SCHREIBER *et al.* (1980) were conducted at 1 atmosphere under anhydrous conditions, fO_2 was controlled. Water or pressure will only affect the Ce oxidation state if they affect fO_2. Whether high fO_2 and low Fe conditions occur during cryptic metasomatism remains to be proven.

The presence of subducted basaltic or crustal components (or derivatives thereof) beneath the OJP has been proposed by BIELSKI-ZYSKIND *et al.* (1984) and NEAL and DAVIDSON (1988). These latter authors have based their conclusions on Sr and Nd isotopic disparities within and between the cpx megacrysts and host alnöite at Malaita, and interpretation of geophysical data (FUROMOTO *et al.*, 1976; HUSSONG *et al.*, 1979). Our present study supports this contention, indicating a proportion of the subducted oceanic material survives to reach the mantle beneath the OJP. The REE signature of the oceanic components has *locally* survived subduction and been incorporated into the mantle, as indicated by CRN211.

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