Reconciling the Shadow of a Subduction Signature with Rift Geochemistry and Tectonic Environment in Eastern Marie Byrd Land, Antarctica.

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Abstract

Basalt-trachyte volcanoes in the Marie Byrd Land (MBL) Cenozoic province lie along the Amundsen Sea coast on the north flank of the West Antarctic rift. Basalts here are characterized by OIB-like geochemistry, restricted ranges of $^{87}\text{Sr}/^{86}\text{Sr}$ (0.702535 – 0.703284) and $^{143}\text{Nd}/^{144}\text{Nd}$ (0.512839–0.513008) and a wide range of $^{206}\text{Pb}/^{204}\text{Pb}$ (19.357–20.934).

Basalts at three MBL volcanoes display two anomalies compared with the above and with all other basalts in West Antarctica. They include $^{143}\text{Nd}/^{144}\text{Nd}$ (0.512778 – 0.512789) values at Mt. Takahe and Mt. Siple that are 2σ lower than other West Antarctic basalts, and Ba/Nb, Ba/La, and Ba/Th values at Mt. Murphy and Mt. Takahe that are 3-8 times higher than normal OIB. Isotope and trace element data do not support crustal and lithospheric mantle contamination, or the presence of residual mantle amphibole or phlogopite as explanations of these anomalies.

The apparent coincidence of these anomalies with the site of a pre-Cenozoic convergence zone along the Gondwanaland margin suggests a subduction influence. Major episodes of subduction and granitic plutonism took place in MBL during the Devonian, Permian, and Late Cretaceous.

Relicts in the source region, of components from these subducted slabs, provide a credible explanation for the uncoupling of Ba from other **large ion lithophile elements (LILE)**, for its erratic distribution, and for the anomalously low $^{143}\text{Nd}/^{144}\text{Nd}$ at Mt. Takahe.

The last episode of subduction ended ~85 Ma, and was followed by continental break-up, rifting and lithospheric attenuation that produced the West Antarctic rift as we know it today. Thus, the enigmatic geochemical signatures in these three volcanoes seem to have been preserved roughly 61-85 m.y. after subduction ended. **New calculations of source melting depth and a new determination of lithospheric thickness suggest that the source of the anomalies resides in a fossil mélange diapir that rose from the Cretaceous subducting slab, became attached to the base of the lithosphere at 80-100 km depth, and remained there during the subsequent plate motion and source remobilization history of this region.**

Keywords. Rift volcanoes. Subduction relicts. Fossil convergent zone. **Mélange diapirs.**

1. Introduction

Previous research has shown that basaltic rocks across the ~3000 km length of West Antarctica (WA) (Fig. 1) display a uniformly narrow range of isotopic and trace element compositions that is unusual for a region this large (Hole and LeMasurier, 1994). For example, $^{143}\text{Nd}/^{144}\text{Nd}$ values range from 0.512839 to 0.513008 and $^{87}\text{Sr}/^{86}\text{Sr}$ from 0.702535 to 0.703520 from the tip of the Antarctic Peninsula to western Marie Byrd Land (MBL), equivalent to the distance from Anchorage to San Francisco (or Stockholm to Seville). A far greater range than this is found among basalts of the western U.S. as a whole, and even along the 800 km length of the Rio Grande Rift. Among trace elements, the uniformly narrow range is not quite so

conspicuous, but still remarkable. Trace element compositions are broadly oceanic island basalt (OIB)-like throughout WA, but high absolute values of Ba and Rb, and $K/Ba < 50$ distinguish MBL basalts from those on the Antarctic Peninsula (Hole and LeMasurier, 1994).

In this context, we describe two volcanoes in eastern MBL, Mt. Takahe and Mt. Murphy (Fig. 1), whose Ba concentrations and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios appear to set them apart from all other basalts in WA. At Mt. Takahe, $^{143}\text{Nd}/^{144}\text{Nd}$ values range from 0.512772 to 0.512790, more than 2σ below all others in WA, except for one at Mt. Siple and one in the Jones Mountains, which lie in the same zone of anomalies as Mt. Takahe and Mt. Murphy, described below. Basalts at both Mt. Takahe and Mt. Murphy are distinguishable from all other WA basalts by Ba/Th, Ba/Nb and Ba/La ratios that lie outside normal OIB ranges. Somewhat more perplexing is the fact that these anomalies lie at the east end of a geochemical gradient that extends across the province, from normal isotopic and trace element compositions in the west (e.g. Hobbs Coast, Fig. 1) to the anomalous values at Mt. Takahe and Mt. Murphy. In following the gradient still further eastward, there is also a pronounced discontinuity in Pb isotopic and trace element compositions that is apparent in crossing from MBL to the Jones Mountains in western Ellsworth Land (Hart et al., 1995).

The west-east gradient may have a paleotectonic significance. Prior to ~85 Ma, there was a convergent plate boundary along the MBL – Zealandia reconstructed continental margin, shown schematically in Fig. 2. The geochemical gradient seems to converge eastward on this fossil convergence zone, suggesting a possible subduction component to the anomalies.

In light of this background, the problems we have addressed in this study are the following. (1) Is the $^{143}\text{Nd}/^{144}\text{Nd}$ anomaly at Mt. Takahe the product of a new source, that would be virtually unique for WA basalts, or is it a result of crustal/lithospheric contamination? (2) The Ba anomaly is especially puzzling, because Ba appears to have been uncoupled from other fluid mobile elements. Is this real? Is it a product of lithospheric contamination, or the result of pre-Cenozoic subduction processes, which Plank and Langmuir (1998) have shown to be capable of uncoupling Ba from other elements? If the latter can be shown to be a reasonable possibility, what are the implications of a subduction signature lingering in the source area ~85 m.y. after subduction ended, and after the subsequent continental break-up, rifting, and lithospheric attenuation that produced the West Antarctic rift as we know it today?

2. Previous Work

The geology, petrology, and history of Mt. Takahe and Mt. Murphy volcanoes, as known before 1987, are described in the “Individual Volcano Description” sections of LeMasurier and Thomson (1990). A thorough study (by Antarctic standards) of Mt. Takahe was conducted in January 1985 when 25 localities were sampled. A total of 24 K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages have now been published, none older than 192 ka (LeMasurier and Rex, 1990; Wilch et al., 1999, 2000).

Previous studies of Mt. Murphy have focused mainly on the glacial history recorded by its morphology, by tillites, pillow lavas and hyaloclastites at and near its base, and by glacial lake beds that are now 1000 m above the present ice level (Andrews and LeMasurier, 1973; LeMasurier et al., 1994; Smellie, 2001). It is also the only MBL volcano exposed, albeit

intermittently, from base to summit and has served as a model of MBL volcano structure. The generalized structure is that of a shield volcano with a basal succession of basalt, probably equal to ~90% of total volume, surmounted by a summit section of felsic and intermediate lavas. Parasitic cones of basanite follow shield building (LeMasurier, 2013). A Quaternary basalt from one of the parasitic cones has been found to be the most primitive basaltic rock in the MBL province, as measured by Mg#, Ni and Cr, and has been used to help model basalt evolution in the Executive Committee Range and the Ames and Flood Ranges (LeMasurier et al., 2003, 2011).

Geochemical studies by Futa and LeMasurier (1983), Hole and LeMasurier (1994), and Hart et al., (1997) form much of the background for this study. Futa and LeMasurier (1983) published the first set of Sr and Nd isotopic data for MBL basalts, which illustrated their OIB affinity. They also proposed the development of pockets in the mantle where metasomatizing fluids enriched the mantle source in large ion lithophile elements (LILE) and light rare earth elements (LREE), while preserving the high $^{143}\text{Nd}/^{144}\text{Nd}$ and low $^{87}\text{Sr}/^{86}\text{Sr}$ inherited from the depleted mantle source. Hole and LeMasurier (1994) presented new REE and trace element data and described the narrow range and lack of temporal change in the compositions of late Miocene and younger basalts throughout WA, suggesting they were derived from an asthenospheric source with no detectable lithospheric contamination. Hart et al., (1997) presented new Pb, Nd, and Sr isotopic data on basalts from the Hobbs Coast (Fig. 1) which demonstrated a HIMU ($^{206}\text{Pb}/^{204}\text{Pb} > 19.8$) signature, but also showed that the full range of basalts lay along a mixing line between HIMU and FOZO (focus zone; $^{206}\text{Pb}/^{204}\text{Pb} = \sim 19.5$) which had depth implications for this region. Modeling, using the algorithms of Herzberg and Zhang

(1996), suggested that FOZO end member basalts were produced by 3.2% melting at 110 km, and the HIMU end members by 1.6% melting at 140 km. Their study provided more evidence for a lack of contamination from ancient continental lithosphere, and noted that the mantle source was once depleted, and subsequently enriched, to account for high concentrations of the Ba to La elements, and negative slope from Nb to Lu, on chondrite-normalized spidergrams.

Wysoczanski et al., (1995), Perinelli, et al., (2011), and Martin et al., (2013) provide data on lower crustal and lithospheric mantle xenoliths from MBL and Victoria Land (Fig. 1). These data are incorporated in our isotope diagrams, and have been used to re-evaluate the possibility that the anomalies at Mt. Takahe and Mt. Murphy were produced by contamination from lower crustal and/or mantle lithospheric sources.

3. Regional Geologic Setting

Unlike the rest of the “Rim of Fire,” the Pacific margin of WA is not a convergent plate boundary, but is a large intra-continental rift system within a plate that has been roughly stationary for the past 85 m.y. (Larter, et al., 2002), except for basin deepening and extension along the western Ross Sea, and possibly within the MBL interior (Cooper, et al., 1991; Cande, et al., 2000; LeMasurier, 2008). Fig. 1 shows the extent and scale of the West Antarctic rift system, and its well-defined southern boundary. In coastal MBL, clearly defined Basin and Range faulting, voluminous alkaline volcanism, and an attenuated crust, which are defining characteristics of the rift, extend to the Pacific coast, and offshore to the Marie Byrd Seamounts, suggesting that the rift has no clear northern boundary (LeMasurier and Landis, 1996; Heinemann, et al., 1999; LeMasurier, 2008).

Fig. 2 is a schematic geological reconstruction that illustrates the convergent plate margin environment that existed in coastal MBL prior to breakup of the New Zealand-Campbell Plateau block (Zealandia) from MBL, and the development of the West Antarctic rift. Forearc basin graywackes are found almost entirely in New Zealand, and the magmatic arc is represented throughout coastal WA by granitic rocks shown in the figure. These are predominantly Late Cretaceous (~90 Ma) peraluminous, calc-alkaline granites, with lesser amounts of Permian and Devonian calc-alkaline granite, representing a subduction environment that characterized this continental margin from Devonian time until subduction ceased at ~85 Ma (Larter et al., 2002); i.e. ~300 m.y. of subduction history with the convergent boundary in roughly the same position. These granitoids intrude lower Paleozoic greenschist to amphibolite facies metaclastic and metavolcanic rocks and granitic orthogneisses (Pankhurst et al., 1998; Mukasa and Dalziel, 2000). Lower crustal rocks are believed to be represented by granulite facies gabbro and norite xenoliths, and the lithospheric mantle by spinel peridotite, wehrlite, websterite and clinopyroxenite xenoliths (Wysoczanski et al., 1995).

Surface wave dispersion measurements by Ritzwoller et al., (2001) suggest a crustal thickness of 25-30 km for coastal MBL. This compares well with a more recent estimate of 22.1-32.8 km by Chaput et al., (2014). Ritzwoller et al., (2001) mapped a prominent asthenospheric anomaly between ~55 km and 200 km throughout the rift system, and they estimate a 50-60 km thickness for the lithosphere in MBL. More recently, however, An et al., (2015) used *S* velocities in the 3-D seismic model AN1-S to calculate upper mantle temperatures of the Antarctic Plate. From these results, they produce maps of the lithosphere-asthenosphere boundary depth on the basis of temperature estimations crossing the 1330°C adiabat and the

uppermost seismic low velocity zone. These maps show a lithospheric thickness of 80-100 km beneath coastal MBL.

The MBL - Zealandia breakup has most recently been estimated at 79.5 – 84 Ma (Wobbe et al., 2012). Pre-breakup rifting is recorded on land in New Zealand back to ca. 100 Ma (Bishop and Laird, 1976; LeMasurier and Landis, 1996) and back to 96 Ma on the sea floor adjacent to MBL (Wobbe et al., 2012). Most of the crustal attenuation in the rift system probably took place during this early rift episode (Lawver and Gahagan, 1994). Somewhat ambiguous sea floor data suggest subduction continued beneath eastern MBL during the interval 84 – 61 Ma, offshore of Mt. Takahe and Mt. Murphy, and is represented in the Chron 27 (61 Ma) reconstruction of Heinemann et al., (1999) by a subduction symbol.

In coastal MBL, the dominant tectonic feature within the rift is the MBL dome – a tectonomagmatic dome defined by elevations of the Late Cretaceous West Antarctic erosion surface (Fig. 1) and the focus of volcanic activity since ~28-25 Ma. Dome-centered patterns of volcanic activity suggest that the dome has risen ~100m/m.y. since the late Oligocene, accompanied throughout its history by basalt volcanism, and joined by felsic volcanism since ~19 Ma (LeMasurier and Rex, 1989; LeMasurier and Landis, 1996). As a result, the oldest felsic volcano is at the crest of the dome (Fig. 1, Mt. Flint), and the youngest, including Mt. Takahe and Mt. Siple, are along its margins. This is similar to the history of the Kenyan and Ethiopian domes in the East African rift (Baker et al., 1972; Rogers, 2006; LeMasurier, 2008).

4. Physical volcanology

Basaltic rocks appear to make up ~90% of the volume of MBL volcanoes. This is a gross estimate, based on studies of four of the 18 MBL shield volcanoes in which thick basal successions of basalt are exposed (LeMasurier, 2013). All but two of the 18 are partially buried beneath the ice sheet, exposing mainly the summit sections, which are composed of felsic and intermediate rocks and basaltic parasitic cones. These exposures create the misleading impression that MBL volcanoes are predominantly felsic. The three volcanoes described below have yielded geochemically anomalous basalt samples, and are the focus of this study.

Mt. Murphy is a late Miocene (9.3-8.2 Ma) basalt shield volcano 32 km in base diameter, with 2200 m of vertical relief (Fig. 3). It is the only MBL volcano exposed from base to summit. A 1400m basal succession of *hy-ol* basalt rests on basement rock, overlain by ~800m of trachyte and benmoreite. Samples from the southwest ridge section are all basalt, from the 500m elevation of the basement contact to the basalt/trachyte contact at 1900m, succeeded by trachyte and benmoreite flows from 1900m to 2446m (LeMasurier et al., 1994). Basanites in Table 1 were found only in parasitic cones. One $^{40}\text{Ar}/^{39}\text{Ar}$ age of 0.592 Ma (Wilch and McIntosh, 2002) is available for this latest stage of activity. Basal succession basalts appear to make up ~90% of the volcano volume (LeMasurier, 2013).

Mt. Siple is quite remote and known only from brief reconnaissance. It rises from below sea level to a 3110m summit. It was first visited in 1983. It consists of a basal succession of *hy-ol* basalt overlain by a felsic summit section. K-Ar ages of 2.0 Ma and <0.1 Ma have been obtained for basal succession sample W83-5C and parasitic cone sample W83-1 (LeMasurier and

Thomson, 1990). Trachytes, collected at the summit during a later visit, yielded an $^{40}\text{Ar}/^{39}\text{Ar}$ age of 0.227 Ma (Wilch et al., 1999). The basalt/felsic contact has not been located.

Mt. Takahe is a late Quaternary (<0.2 Ma) trachytic shield volcano ~32 km in base diameter, with 2000 m of relief above ice level (Fig. 4). It lies 200 km inland from the Amundsen Sea coast, partly buried by the West Antarctic ice sheet, which is >2 km thick at this locality (Drewry, 1983). Based on comparisons with Mt. Murphy, Mt. Siple, and two other more fully exposed volcanoes, it seems likely that the trachyte is a summit section that rests on a much larger basaltic base (LeMasurier, 2013). The exposed basaltic rocks at Mt. Takahe are basanites and hawaiites, found in parasitic cones that probably make up <1% of the exposed volume. In more fully exposed volcanoes parasitic cone basalts are geochemically similar to basal successions at Mt. Bursey and the Crary Mountains (Fig. 1), but are more silica undersaturated than the basalts exposed at the base of Mt. Murphy and Mt. Siple. The 10 analyses in Table 1 represent all accessible exposures of basaltic rock on Mt. Takahe.

5. Geochemistry

5.1 Analytical methods

The polar environment provides exceptionally fresh, unweathered, and unoxidized samples for analysis. This is reflected in low H_2O , CO_2 , and $\text{Fe}_2\text{O}_3/\text{FeO}$ of samples in Table 1, which are the original raw data. Glasses are fresh and unhydrated. Loss of mobile elements from devitrification is unlikely. Samples were selected for this study based on their classification as basanites, basalts and hawaiites (Fig. 5A).

Ten new major element, 21 new trace element, and 4 new isotope analyses are presented in Table 1. Major elements and originally all trace element analyses were performed by X-ray fluorescence (XRF) at the University of Nottingham (UK), except for samples 67-67A and 67-65B, which were carried out at the Washington State University Geoanalytical Lab; FeO was determined by potassium dichromate titration and F by selective ion electrode. Subsequently, five Mt. Takahe basalt samples were selected for inductively coupled mass spectrometry (ICPMS) re-analysis of 38 trace elements at the Open University (UK), following procedures described by Rogers (2006). Accuracy is generally better than 10% and often significantly better than 5%, although accuracy for a few elements lies outside these limits. Precision at the 1 sigma level is ($\pm 2\%$). Analysis of the standard JB-2 is included in Table 1S for inter-lab comparison with the Göttingen analyses described below.

New ICPMS trace element analyses (36 elements in Table 1) of eight Mt. Murphy samples and two Mt. Siple samples, were performed at Göttingen University. Here, 120 mg of powdered sample was reacted with 3 ml of ultra-pure HF/HClO₄-HF/HNO₃ mixture for 12 hours at 180°C in pressurized teflon beakers using the DAS30 Picotrace^R system at Geowissenschaftliche Zentrum University of Göttingen. Solutions were evaporated to dryness for 2 hours at 150°C. The precipitates were dissolved in 2 ml ultra-pure HNO₃, and 200 µl of a 10 ppm solution of In and Re as internal standards was added. Blanks were prepared simultaneously. The standard material JA-2 was dissolved and analyzed together with the samples, and from standard measurements the 2 σ error of the method is estimated to be <20% for Nb and Ta, <10% for Be, Cs, Cu, Hf, Li, Y, Pb, Rb, Tl, Th, and U, and <10% for the rare earth

elements (REE) except for Ho, Er, Tm, Lu (<20%). ICPMS solution measurements were made on a FISIONS VG PQ STE ICPMS instrument.

The four Pb isotope analyses in Table 1, and accompanying Sr and Nd isotope analyses were carried out at Woods Hole Oceanographic Institution (Mass., USA) with conventional techniques (Hauri and Hart, 1993), using sample powders which had been leached in warm 6 N HCL for 1 hour. Sr and Nd data carry precisions of 0.003%, and are reported relative to 0.71024 (NBS 987) and 0.511847 (La Jolla). Pb is corrected ratio by ratio for fractionation, using NBS 981 and the values given by Todt et al., (1996).

Sr and Nd isotope analysis of **Mt.** Takahe sample 85-19, including chemical separation and measurement was performed at the University of Michigan using VG Sector thermal ionization mass spectrometers. Analytical procedures are given in Mukasa et al., (1987, 1991). $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were corrected for instrumental mass fractionation by normalizing to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ and $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$, respectively. Replicate analyses of NBS-987 and La Jolla Nd standards gave $^{87}\text{Sr}/^{86}\text{Sr} = 0.710254 \pm 0.000011$ ($N = 30$, $2\sigma_m$) and $^{143}\text{Nd}/^{144}\text{Nd} = 0.511850 \pm 0.000019$ ($N = 30$, $2\sigma_m$).

6. Rock Compositions

6.1 Major elements

Major element data are presented in Table 1. In addition to **Mt.** Takahe and **Mt.** Murphy data, most of the diagrams that follow include data from 15 **additional** MBL localities and 6 from the Jones Mountains (Hart et al., 1995), hereafter referred to as the comparative suite

(Fig. 1 shows all localities). Geochemical characteristics of Mt. Takahe and Mt. Murphy are most distinctive and most easily interpreted in the context of the entire province. This kind of presentation also helps us evaluate whether their characteristics represent normal OIB variation, or involve other factors. Jones Mountains basalts have been well studied, and display trace element anomalies that are similar to those at Mt. Takahe and Mt. Murphy in many respects (Hart et al., 1995).

Fig. 5A presents the classification of Mt. Takahe and Mt. Murphy basaltic rocks following LeBas et al., (1986). Those at Mt. Takahe are all *ne*-normative, including basanites (*ol* >10%) and hawaiites. At Mt. Murphy, the basal succession consists of *hy-ol*- normative subalkaline basalts and alkaline basalts (*ne* = 1.04 – 1.88); parasitic cones are alkaline basalts somewhat more undersaturated than the basal succession (*ne* = 4.04 – 9.07). From only two analyses, Mt. Siple seems to resemble Mt. Murphy in that the basal succession sample is *hy-ol*- subalkaline basalt and the parasitic cone is *ne*-hawaiite (Table 1). Compared with Hawaiian basalts, all MBL samples would be considered alkali basalts (Macdonald and Katsura, 1964). Fig. 5B shows that Mt. Murphy and Mt. Siple basal successions are unusual for this province in their relatively less alkaline character, joined by only two Hobbs Coast samples, one of which is a *hy-ol* basalt at Patton Bluff (Hart et al., 1997).

Fig. 6A shows that Mt. Takahe and Mt. Murphy basaltic rocks fall into three groups. All Mt. Takahe rocks, and Mt. Murphy parasitic cones, form a high soda group, along with most other MBL basaltic rocks. The Mt. Murphy basal succession and Patton Bluff *hy-ol* basalt form a less sodic group. In addition, Mt. Murphy parasitic cones are more MgO-rich than all Mt.

Takahe basanites and most basal succession basalts. The two *Mt. Siple* samples are intermediate in their Na₂O and MgO character.

Fig. 6B shows that the high MgO of *Mt. Murphy* parasitic cones and low MgO of *Mt. Takahe* basanites and *Mt. Murphy* basal succession basalts correlate closely with Ni.

Fig. 6C shows that *Mt. Murphy* basal successions are generally more CaO-rich than other MBL basaltic rocks. Fig. 6D plots CaO/Na₂O vs. V, both of which are sensitive to clinopyroxene (cpx) fractionation. The CaO-rich rocks can be seen to be rich in V, and have a high CaO/Na₂O ratio, suggesting that *Mt. Murphy* basal succession basalts have experienced less cpx fractionation than the other MBL rocks.

6.2 Trace elements

Trace element data are presented in Table 1 and Figs. 7, 8, 9, and 12. All diagrams that include La, Ce, Th, or Pb use ICPMS data only.

The general trace element characteristics of *Mt. Takahe* and *Mt. Murphy* are illustrated in Fig. 7 by seven samples. The most distinguishing and anomalous element is Ba. There are three high Ba and nine low-Ba samples at *Mt. Murphy*. The high-Ba samples are *hyol* basalts from localities 32 and 33 (Fig. 4; Table 1S). Low-Ba samples 32E and 33F are from the same sample sites. To illustrate the range of trace element characteristics, Fig. 7 includes hi-Ba and lo-Ba samples from *Mt. Takahe* (67-66B, 85-19) and *Mt. Murphy* (85-32B, 85-31A, 67-62), together with a representative sample from the Ames and Flood Range (AR/FR) group, which

296 modeling results suggest contain less than 1% crustal contamination (LeMasurier et al., 2011).
297 Also included is a representative sample from the Jones Mountains (Hart et al., 1995).

298 The Ta-Nb hump and negative spikes for K and Pb are familiar OIB characteristics (Hart
299 and Gaetani, 2006). However, both Ba and Pb characteristics are erratic, as seen by the
300 following. The high Ba samples do not show a commensurate enrichment in other LIL elements
301 (e.g. Rb, Th, K); but show a more negative dip in Rb and Th than the more normal AR/FR
302 sample. **Mt.** Takahe sample 67-66B has the highest Pb of all seven samples and also a high Ba
303 spike; **whereas** sample 19 has a moderately high Ba spike, but a large negative Pb spike. Both
304 the high-Ba and low-Ba **Mt.** Murphy samples display large negative Pb spikes. The Jones
305 **Mountains** sample shows low-Ba, but high-Pb. **Mt.** Siple and Patton Bluff *hy-ol* basalts (not
306 shown) show neither Ba nor Pb enrichment.

307 Binary trace element plots are presented in Figs. 8A-F. The major observations apparent
308 from these diagrams are (1) Ba is largely uncoupled from other LILE, (2) MBL basalts have close
309 affinities to OIB, but with some significant deviations, and (3) there is a pronounced gradient in
310 some trace element characteristics, from west to east, across the MBL dome among near-
311 coastal volcanoes. The data used to position OIB mantle end members in these diagrams are
312 found in spidergrams in Hart and Gaetani (2006), and in Tables 1S and 2S, which list data that
313 are as yet unpublished.

314 Fig. 8A shows that most **Mt.** Takahe samples are richer in Ba than all other MBL basalts
315 except the three hi-Ba samples from **Mt** Murphy, which are in a class by themselves. Conversely
316 all the **Mt.** Murphy parasitic cone basalts and the lo-Ba basal succession basalts are among the

317 lowest Ba basaltic rocks in the province. An eastward decrease in Nb can also be seen here, and
318 in Figs. 8B, 8D, and 8F.

319 Fig. 8B is intended to help visualize a contrast among the highly incompatible elements
320 K, Nb, and Ba. In contrast to Fig. 8A, most Mt. Murphy samples, including the three high-Ba
321 samples, plot among all other MBL basalts suggesting that Ba is uncoupled from the other LILE.
322 Similarly, while most Mt. Takahe basaltic rocks are richer in Ba than all other MBL basalts, only
323 two are slightly higher in K than all other MBL basalts. Fig 8C provides another example in which
324 Ba distinguishes Mt. Takahe and Mt. Murphy basalts from others in the region.

325 Figure 8D is taken, in modified form, from Hole and LeMasurier (1994), with data for
326 HIMU, EM1, EM2, and N-MORB end members from Hart and Gaetani (2006) and Table 1S.
327 Ratios of these highly incompatible elements should not change during fractional crystallization
328 or partial melting, and are likely to represent source characteristics. The plot shows that MBL
329 basalts lie generally within the domain of OIB mantle end members, with the exception of the
330 three high-Ba Mt. Murphy and most Mt. Takahe samples. All but one of the Mt. Takahe rocks
331 are separate from the other MBL basalts and appear mildly shifted toward high Ba/Nb, higher
332 in Ba/Nb than end member OIBs. A west to east gradient can be seen in this diagram too, in
333 which Hobbs Coast and AR/FR volcanoes trend toward a HIMU-rich source, and Mt. Takahe to
334 an EM-rich source.

335 The fields for Antarctic Peninsula and Hudson Mountains basalts, from Hole and
336 LeMasurier (1994), show that the Mt. Takahe/Mt. Murphy Ba anomaly is truly unusual among

all West Antarctic basalts, except for those in the Hudson Mtns, which lie east of Mt. Murphy and west of the Jones Mountains (Fig. 1).

Fig. 8E presents two ratios that constrain the OIB affinities of MBL basalts. In OIB, Ba/La varies within the range 6-9, and Ba/Nb within 4-9 (Table 2). All Mt. Takahe samples lie outside these ranges, as do the three high-Ba Mt. Murphy basalts, and most AR/FR and Jones Mountains basalts. All but one Hobbs Coast and four low-Ba Mt. Murphy basalts lie within OIB boundaries. One low-Ba Mt. Murphy basalt lies barely outside these boundaries.

Fig. 8F plots two HFSE that are immobile in aqueous fluids. All MBL samples, including high-Ba and low-Ba samples, form a coherent linear trend that illustrates a constant Nb/Ta ratio for the province. It shows again the systematic eastward decrease in Nb, and also in Ta.

Two Fig. 9 diagrams explore the contrast between MBL and Jones Mountains basalts with respect to Ba and Pb. A Ce/Pb ratio >25 is a widely cited value for uncontaminated OIB (Hofmann et al., 1986); those with lower values are usually interpreted to involve crustal contamination (Hart et al., 1995). In Fig. 9A we have inverted this ratio for ease of visualization, so that indicators of a subduction influence (Ba/Nb and Pb/Ce) will rise to the right in concert. On this plot, all but one of the Mt. Takahe basaltic rocks has a Pb/Ce ratio <0.04 (equivalent to Ce/Pb>25), which implies normal OIB, but this is contradicted by Ba/Nb >9, seen here and in Fig. 8E.

All Mt. Murphy basalts lie below Pb/Ce = 0.04, along with most of the comparative suite. For Jones Mountains basalts and the one hi- Pb Mt. Takahe sample, Pb/Ce > 0.04. Thus, for Mt. Murphy, only Ba/Nb and Ba/La for the high-Ba basalts (Figs. 8C, 8D) imply significant departures

from OIB. This high Ba characteristic is shared by all Mt. Takahe basalts. By contrast, Jones Mountains basalts are the only ones that combine high Pb/Ce, with low Ba/Nb (Fig. 9A). Basalts that show the least deviation from uncontaminated OIB, in Figs. 8 and 9, are those from the Hobbs Coast, AR/FR, and Crary Mountains.

Fig. 9B shows a great deal of scatter, but is included here to emphasize the difference between MBL and Jones Mountains basalts. The plot shows significantly high Pb for Jones Mountains and the high-Pb Mt. Takahe basalt, consistent with the relations in Fig. 9A.

Fig. 9C evaluates whether variation in percent partial melting could be a factor in trace element patterns. Inspection of Table 1 shows that Nb increases with increasing silica undersaturation, i.e. with increasing normative *ne*. Thus the Nb/Y ratio has been used as an index of percent partial melting (Hole, 1988), i.e. the percent partial melting decreases with increasing Nb/Y. In this figure, the easternmost volcanic centers (Mt. Takahe, Mt. Murphy, Jones Mountains) lie at the low Nb/Y end of the array, and the westernmost centers, Hobbs Coast and AR/FR lie at the middle and high Nb/Y end. Of particular interest is the Hobbs Coast array, because the depths and percent partial melting that produced these basalts have been modeled by Hart et al., (1997). Their results show that basalts at Coleman Nunatak (Fig. 1) were formed by an estimated 1.6% melting at 140km depth, whereas the remaining basalts formed at progressively shallower depths and increasing percent melting, up to ~3.2% melting at 110 km for those at Patton and Holmes Bluffs (separate circle). The pattern in Fig. 9C conforms nicely with these conclusions, in that the Coleman basalts have the highest Nb/Y and the Holmes and Patton basalts have the lowest, among Hobbs Coast basalts.

Table 2 summarizes trace element ratios for five MBL volcanic ranges, which include 13 major shield volcanoes, and compares them with the ratios for normal OIB. At Mt. Takahe and Mt. Murphy, Ba/Nb, Ba/Th, and Ba/La ratios are 2-5 times higher than normal OIB. Ba/Th remains above normal OIB for all other MBL volcanoes, whereas Ba/Nb, Ba/La, and K/Nb are within normal OIB ranges.

7. Isotopes

Sr, Nd, and Pb isotope data for Mt. Takahe, Mt. Murphy, and Mt. Siple basaltic rocks are presented in Table 1 and Figs. 10 and 11. Although Mt. Siple trace element compositions are within the normal range for OIB (except Ba/Th), it has isotope ratios that are anomalous compared with basaltic rocks elsewhere in WA.

Fig. 10A presents Sr and Nd isotopic data in relation to the fields for MORB, HIMU, EM1, EM2 and potential MBL upper crust, lower crust, and lithospheric mantle contaminants (see Fig. 10 and 11 references). Some of the more evolved rocks are plotted here to show that Mt. Takahe basaltic rocks have more enriched Sr and Nd isotopic compositions than the mugearites, benmoreites and trachytes. The Mt. Takahe basalts cluster at the lowest $^{143}\text{Nd}/^{144}\text{Nd}$, highest $^{87}\text{Sr}/^{86}\text{Sr}$ end of the array, with $^{143}\text{Nd}/^{144}\text{Nd}$ ratios at least 2σ lower than all others, except for the Mt. Siple parasitic cone basalt and one Jones Mountains sample (see also Fig. 10C). For the province as a whole, the low $^{143}\text{Nd}/^{144}\text{Nd}$ and high $^{87}\text{Sr}/^{86}\text{Sr}$ of Mt. Takahe and Mt. Siple significantly extend the ranges of these ratios toward the EM1 end member component.

Fig. 10B, from Hole and LeMasurier (1994), illustrates the unusually narrow range for $^{143}\text{Nd}/^{144}\text{Nd}$ among basalts in WA compared with basalts from the spatially and tectonically

analogous western U.S. Subsequent (to 1994) data from MBL and Ellsworth Land do not change this picture significantly, except that the $^{143}\text{Nd}/^{144}\text{Nd}$ values reported here for Mt. Takahe basalts are 2σ lower than all other basalts in WA for which we have data. Thus, although the Mt. Takahe basalts (plus one Mt. Siple and one Jones Mountains) all lie within the $^{143}\text{Nd}/^{144}\text{Nd}$ range for OIB (Table 1S), and in that context are unremarkable, they appear to be significantly, perhaps uniquely, anomalous among basalts within the 3000 km span of WA.

Fig. 10C plots $^{143}\text{Nd}/^{144}\text{Nd}$ vs $^{87}\text{Sr}/^{86}\text{Sr}$ with the geographic localities of the source volcanoes identified (cf. Fig. 1). This plot shows a west to east gradient in isotopic composition among near coastal volcanoes. It includes data from Mt. Takahe, Mt. Murphy, and Mt. Siple, plus the comparative suite, and shows that the far western Hobbs Coast and AR/FR rocks cluster at the high $^{143}\text{Nd}/^{144}\text{Nd}$, low $^{87}\text{Sr}/^{86}\text{Sr}$ end of the array. The Mt. Murphy basalts plot at lower Nd and higher Sr isotope ratios, along with samples from the MBL dome crest (Mt. Flint and Mt. Petras) and most Jones Mountains samples. Mt. Takahe basalts cluster at the lowest $^{143}\text{Nd}/^{144}\text{Nd}$, highest $^{87}\text{Sr}/^{86}\text{Sr}$ end of the array, along with one Mt. Siple and one Jones Mountains basalt. The inland and eastern Executive Committee Range (ECR) and Crary Mountains basalts cluster with far western coastal volcanoes and do not fit the dominant pattern.

Fig. 11 shows that the full range of MBL Pb data forms an array between MORB and HIMU, with little overlap onto the fields of potential basement contaminants. It also shows how Pb data for Mt. Takahe, Mt. Murphy, and Mt. Siple compare with other volcanic centers in the province, and with the Jones Mountains. The full range of $^{206}\text{Pb}/^{204}\text{Pb}$ in MBL is 19.36 – 20.69.

Mt. Takahe, Mt. Murphy and Mt. Siple data lie well within this range (19.67-20.24). However, the $^{208}\text{Pb}/^{204}\text{Pb}$ (39.25-39.99) values extend the range toward the EM2 component. AR/FR basalts are mostly grouped on the low $^{207}\text{Pb}/^{204}\text{Pb}$ and low $^{208}\text{Pb}/^{204}\text{Pb}$ sides of the two arrays, and Mt. Takahe, Mt. Murphy and Mt. Siple basalts mainly on the high sides. Jones Mountains basalts lie nearer EM1 and further from HIMU end members than all MBL basalts.

8. Discussion

Major element chemistry divides Mt. Takahe, Mt. Murphy, and Mt. Siple basaltic rocks into three groups, based largely on silica saturation characteristics and percent MgO. The most likely origins of these groups do not seem relevant to understanding trace element and isotopic anomalies. It seems likely, for example, that the *hy-ol* basalts at Mt. Murphy and Mt. Siple are products of a larger percent melting, at shallower depths, than the source of the more undersaturated alkali basalts, as has been modeled for the Patton Bluff *hy-ol* basalt (Hart, et al., 1997). Most of the remaining major element characteristics of the three groups can be reasonably explained by variable fractionation of olivine, suggested by the close correlation between Ni and MgO (Fig. 6B), and to a lesser extent, clinopyroxene (Fig. 6D). Deviations from OIB and from other WA basalts are recognizable only from trace element and isotope data. The key issue is to determine whether these anomalies are related to contamination, or to source characteristics.

8.1 The Ba anomaly

The concentration of Ba, and the ratios Ba/Th, Ba/Nb, and Ba/La are 2-3 times higher than OIB for all but one Mt. Takahe basalt. Ba ratios among Mt. Murphy basalts are more

unusual, in that three samples are 3-8 times higher than OIB, and the remainder are among the lowest in MBL. Furthermore, both high-Ba and low-Ba samples were collected from two of the same localities (localities 32 and 33; Fig. 3, Table 1). In what follows, we attempt to test whether these values are primary magmatic characteristics, or products of post-eruptive contamination.

In addition to the usual precautions taken to collect fresh, unaltered samples and avoid lab contamination, we searched for evidence of barite contamination and could find none. Barite is a low temperature hydrothermal mineral. Its deposition would most likely be accompanied by hydrous alteration and propylitization, the latter characterized by epidote, chlorite, carbonates and albitization of feldspars. Propylitization would be obvious in the field and in thin section, and together with the negligible H₂O and CO₂ in the analyses (Table 1), we find no reason to consider the Ba anomalies as anything other than real magmatic characteristics.

The fields for potential crustal and mantle contaminants are shown in Figs. 10a and 11. The data base for the upper crust field (MBL granitoid and gneiss) is based on studies of exposed rocks (Pankhurst et al., 1998; Mukasa and Dalziel, 2000) and is quite robust. The data for lower crust and mantle fields are derived from xenolith studies and are of questionable value for interpreting source characteristics, i.e. only one sample from the cited xenolith suites contains phlogopite, the occurrence of which would strongly suggest a lithospheric mantle source. However, this sample is from Mt. Morning (Fig. 1), 1800 km west of Mt. Takahe (Martin, et al., 2013).

Contamination from the upper crust should be readily seen in isotopic compositions. The fact that Mt. Takahe basaltic rocks have more enriched Sr and Nd isotopic compositions than the mugearites, benmoreites and trachytes (Fig. 10A), and the Nb-Ta hump is within the normal range for OIB (Fig. 7), argues against basement contamination.

Mafic granulite xenoliths from the ECR (Fig. 1) (Wysoczanski et al., 1995) are believed to represent the MBL lower crust. They have more enriched Sr isotopic compositions and elevated $^{143}\text{Nd}/^{144}\text{Nd}$ ratios at a given $^{87}\text{Sr}/^{86}\text{Sr}$, as well as less radiogenic Pb isotopic compositions, compared with basalts from Mt. Takahe and Mt. Murphy.

Spinal peridotite xenoliths from Mt. Morning (Fig. 1) have been interpreted to represent lithospheric mantle (Martin et al., 2013). The available data show higher $^{143}\text{Nd}/^{144}\text{Nd}$ and less radiogenic Pb isotopic compositions than Mt. Takahe or Mt. Murphy basaltic rocks (Figs. 10a and 11). Pyroxenite xenoliths from the ECR, Mt. Morning, and Northern Victoria Land (Fig. 1) range from wehrlite to clinopyroxenite to websterite (Wysoczanski et al., 1995; Perinelli et al., 2011; Martin et al., 2013). Pyroxene-rich rocks have greater melt productivity than peridotite because of a much lower solidus (e.g. Hirschman and Stolper, 1996) and therefore could represent an enriched mantle component. Like the peridotite xenoliths, however, the pyroxenite xenoliths have elevated $^{143}\text{Nd}/^{144}\text{Nd}$ ratios at a given $^{87}\text{Sr}/^{86}\text{Sr}$, and less radiogenic Pb isotopic compositions than Mt. Takahe and Mt. Murphy basalts (Figs. 10 and 11). None of the ultramafic or lower crustal xenoliths have characteristics that constrain source depth, or that might represent contaminants that could explain the Ba and Nd isotope anomalies.

Hart et al. (1997) discuss the possibility that phlogopite and/or amphibole in the mantle lithosphere could account for the high Ba/Th of Hobbs Coast basalts, but found no clear evidence that this could be effective, without producing anomalies in other trace element ratios which we do not see. However, amphibole is a likely residual phase in the Hobbs Coast source (Hart, et al., 1997) and we have examined the possibility that pargasite in mantle lherzolite might have been a factor in producing the eastern MBL Ba anomaly. The Ba K_D for pargasite in garnet peridotite, in equilibrium with basanite liquid, is similar to that for Rb and much higher than that for Th (Adam and Green, 2011), suggesting that Th should be enriched well above Ba, and Rb enrichment should be similar to that for Ba. We do not see this in the Fig. 7 spidergrams, especially among the Ba-rich samples, i.e. there is no evidence here for an enriched lithospheric mantle source.

Finally, we observe that the east end of the trace element and isotopic gradient appears to approach the pre-Cenozoic convergence zone that existed for 300 m.y. along the reconstructed MBL-Zealandia continental margin (Fig. 2). This suggests that a subduction influence on mantle source composition should be considered. Fig. 2 is schematic, and the intersection of the gradient with the fossil convergence zone is speculative, but a similar position for the convergence zone is shown in the recently published reconstruction of Yakymchuk et al., (2015), i.e. the geographic aspect of this possibility is not unreasonable. Furthermore, Engebretson et al. (1992) estimate that up to 6 km² of lithosphere per km² of Earth surface area was subducted beneath MBL in just the past 180 m.y., which suggests that a robust source for these anomalies was at one time available.

The high Ba spike relative to other LILE (Fig. 7), and Ba/Nb >9, can be interpreted as a component of subducted sediment in the mantle source, especially in the context of the HIMU OIB signature, common to all the MBL rocks, the origin of which has been ascribed to removal of hydrous fluid and LILE from subducted slabs (Weaver, 1991). The influence of aqueous fluids is also suggested by the less coherent variation of LILE compared with HFSE (e.g. Fig. 8F vs 8B), in addition to the previously described evidence for mantle metasomatism (Futa and LeMasurier, 1983; Hart et al., 1997).

Fig. 12 compares Mt. Takahe and Mt. Murphy samples with the Ames Range, GLOSS (Plank and Langmuir, 1998) and average EM1 (Hart and Gaetani, 2006). The EM1 comparison is consistent with Figs. 8D and 10C in suggesting that EM1 is an important component in the source of the Ba-rich basalts at Mt. Takahe and Mt. Murphy. The GLOSS comparison suggests that a component of subducted sediment in the mantle source could be the source of high Ba in Mt. Murphy and Mt. Takahe basalts, and of the elevated Pb of one Mt. Takahe sample.

Plank and Langmuir (1998) provide a mechanism for uncoupling Ba from other LILE and for explaining the erratic nature of Ba distribution. They note that inconsistencies in Ba distribution can be attributed to inhomogeneities within the slab source, and also to the erratic nature of Ba transfer and subsequent emplacement in the overlying mantle wedge. For example, some diatomaceous and radiolarian oozes are rich in Ba, but some are not, and hydrothermal fluxes in the trench environment may, or may not, lead to Ba enrichment. They explain that this complex behavior leads to erratic distribution and redistribution of Ba in the subduction environment and to the uncoupling of Ba from other elements. Similarly, it has

been proposed that the distribution of Pb in oceanic crust can be influenced by hydrothermal processes (Peucker-Ehrenbrink et al., 1994; Hart et al., 1997). These relationships provide a reasonable explanation for the erratic characteristics of the Ba anomalies, via hydrous fluids removed from the slab.

8.2 The $^{143}\text{Nd}/^{144}\text{Nd}$ anomaly at Mt. Takahe

A comparison of Figs. 8D and 10C relates isotopic characteristics to trace element source indicators. Figs. 10A, 10C and 8D all suggest that the Mt. Takahe source is relatively rich in EM end member OIB components, whereas the Hobbs Coast and the AR/FR sources are more HIMU-rich. The slab mélange includes pelagic and terrigenous sediments that are believed to make up ~3% of the EM source (Weaver, 1991). As the term implies, they are intensely mixed along the interface between the slab and adjacent mantle wedge (e.g. Marschall and Schumacher, 2012; Vogt and Gerya, 2014). Incorporation of EM1 component, with its very low $^{143}\text{Nd}/^{144}\text{Nd}$ (Fig. 10A), provides a likely source for the Takahe isotope anomaly.

A comparison of Fig. 10C with Fig. 1 shows that the gradient toward increasing EM source components takes place generally eastward across the MBL dome. However, if the far inland ECR and Crary Mountain samples are removed, the gradient is more convincing, i.e. the ECR points should lie with the dome crest basalts, and the Crary Mtns points should lie near Takahe, based on geographic position alone (Fig. 1). This suggests that there is a coastward component as well as an eastward component to the gradient. A coastward component is to be expected if the source of the pattern is related to subduction, since Stern, et al. (1990) found that slab components in young Andean lavas decrease eastward, away from the arc. If this

eastward and coastward increase in EM components is valid, then the trend follows the eastward termination of subduction along the Pacific coast of Antarctica in Late Cretaceous time (Larter, et al., 2002). Conversely, the westward decrease of EM components toward the Hobbs Coast suggests that the zone of geochemical anomalies lies further offshore at the Hobbs Coast, and the coastline converges eastward on the zone of anomalies, just beginning to intersect them at Mt. Siple, and more centered over them at Mounts Takahe and Murphy.

In addition to the eastward gradient in Nd and Sr isotopic compositions, there is an eastward and coastward decrease in Nb (Fig. 8F). Fig. 9C (Nb/Y vs. Nb) shows an accompanying decrease in Nb/Y, which implies a general eastward increase in percent partial melting. This is supported by the gradient among Hobbs Coast samples for which source depth and percent partial melting have been modeled (Hart, et al., 1997), as described above. To extend this work to the present study, pressure and temperature estimates of melting for the five highest Mg# samples from Mt. Murphy were determined using the model of Lee et al. (2009). The five samples form a tight cluster between 1474 - 1494°C and 2.7-3.0 GPa (80-90 km depth). The lowest pressure sample is an older shield sample, whereas the other four are from younger parasitic cones (Fig. 3; Table 3S). This suggests a fairly consistent melting regime over much of the ~9 m.y. history of Mt. Murphy. It appears, therefore, that the anomaly gradient is related to an eastward increase in EM components, an increase in percent partial melting, and a decrease in source depth, all of which accompany the tapping of a source that contains progressively more remnants of the pre-Cenozoic subduction mélange.

There is a remarkable coincidence between this new calculation of source melting depth, at 80-90 km, and the newly published seismic data (An et al., 2015) which indicate an 80-100 km lithospheric thickness beneath MBL. These suggest that the Mt. Murphy and Mt. Takahe source lay at the base of the lithosphere. This source region was apparently impregnated by aqueous fluids and sedimentary components derived from Cretaceous, or conceivably older, slabs.

If this interpretation is approximately correct, then these geochemical relicts have persisted in the mantle source for as much as 85 m.y. after subduction ended. Although it is well established that de-watering of subducted slabs metasomatizes the overlying mantle wedge, there is still uncertainty about how long the signature of this process remains in the mantle wedge after subduction ends. Farmer et al., (2013) found that the metasomatized mantle wedge in the northern Sierra Nevada (California) was not involved in any post-subduction magmatism. At the other extreme, Pegram (1990) argued that the Mesozoic rift tholeiites of the eastern U.S. were derived from mantle that was impacted by Grenvillian (1 Ga) subduction. Closer to the subject of this paper, late Miocene basalts in the Jones Mountains display low Ce/Pb ratios (8.5-17.6), low $^{207}\text{Pb}/^{204}\text{Pb}$, and high Pb/Pb*, that have been interpreted by Hart et al., (1995) as the mark of subduction impregnated subcontinental lithosphere that has remained in the source area roughly 50-60 m.y. after subduction ended in this region. Thus, the suggestion that a subduction signature remains ~85 m.y. after subduction ended in MBL is not without precedent.

587 8.3 The Ba-Pb discontinuity

588 Although some aspects of the west – east geochemical gradient continue smoothly from
589 Mt. Murphy to the Jones Mountains (e.g. Nb), there is an abrupt discontinuity in Ba and Pb
590 trace element and isotopic anomalies between the two. High Ba/Nb, and Ba/La distinguish Mt.
591 Takahe and Mt. Murphy basalts from the Jones Mountains and the rest of MBL (Table 2, Figs.
592 8C, D, E), whereas high Pb/Ce and low Th/Pb separate the Jones Mountains from all MBL (Figs.
593 9B, D). Similarly, low $^{143}\text{Nd}/^{144}\text{Nd}$, coupled with low $^{87}\text{Sr}/^{86}\text{Sr}$ at a given $^{143}\text{Nd}/^{144}\text{Nd}$ (Fig. 10B),
594 distinguishes Mt. Takahe from the Jones Mountains and all other MBL volcanoes, whereas low
595 $^{206}\text{Pb}/^{204}\text{Pb}$ separates the Jones Mountains from all MBL (Fig. 11).

596 The area of this discontinuity ($\sim 100^\circ$ W) lies near the Late Cretaceous/ early Cenozoic
597 boundary between the Antarctic plate and the Phoenix plate (Eagles et al., 2004 ; Larter et al.,
598 2002), which was a complex zone of rapidly changing plate boundaries at that time. Larter et
599 al., (2002) show that subduction continued from Thurston Island (Fig. 1) eastward at least until
600 45 Ma, long after it had stopped farther to the west. Plank and Langmuir (1998) have found
601 that there is no common stratigraphy among trench sections. Each subduction environment is
602 characterized by a different sedimentary package. This must certainly have been the case
603 where subduction ended adjacent to MBL at ~ 85 Ma, but continued perhaps as much as 40 m.y.
604 longer off the coast of the Jones Mountains. We have no data on the composition of trench
605 sediments in either of these environments, but it seems reasonable to speculate that they were
606 different from each other, and that this resulted in different subduction signatures for MBL
607 compared to the Jones Mountains.

8.4 Implications for mantle dynamics

The preceding discussion has led us to propose that processes operating within a subduction environment provide the most reasonable explanations for (1) the high – Ba anomalies, (2) their uncoupling from other LILE, (3) the anomalously low $^{143}\text{Nd}/^{144}\text{Nd}$ at Mt. Takahe, and (4) the eastward and coastward gradient of anomalies across the province. In this context, the Marschall and Schumacher (2012) mélange–diaper model of the subduction environment provides a unifying conceptual framework that embraces the processes interpreted above.

The key elements to the model involve the development of instabilities along the slab mantle interface during subduction, and consequent rise of mélange diapirs into the mantle wedge. During their rise, these materials are heated, cross the dehydration curves for serpentine and chlorite yielding hydrous fluids, and are partially melted yielding arc volcanism and, in MBL, the Cretaceous calc-alkaline granitoids.

The new lithospheric thickness data, the calculated depth of melting beneath Mt. Murphy, and the anomalies themselves, suggest that the remains of such a diaper are now attached to the base of the MBL lithosphere and were remobilized during late Cenozoic volcanism. The long term preservation of the subduction signatures (i.e. the anomalies) suggests that the source materials remained attached to the base of the lithosphere during any subsequent motion of the plate.

9. Summary and Conclusions

1. Basaltic rocks at Mt. Takahe and Mt. Murphy volcanoes display anomalies in Ba/Nb, Ba/La and Ba/Th that are well outside normal OIB ranges, and outside the ranges for basaltic rocks anywhere else in West Antarctica, except the Hudson Mtns.

2. Basaltic rocks at Mt. Takahe and Mt. Siple volcanoes display $^{143}\text{Nd}/^{144}\text{Nd}$ ratios that, although within the range for normal OIB, are 2σ below all other basaltic rocks in WA.

3. Isotopic and trace element data do not support crustal or lithospheric mantle contamination, or residual phlogopite or pargasite in the mantle source, as viable explanations for these anomalies. No other probable high-Ba, low radiogenic Nd lithospheric source has yet been reported from MBL.

4. There is a geochemical gradient in MBL, characterized by eastward and coastward decreasing $^{143}\text{Nd}/^{144}\text{Nd}$, increasing $^{87}\text{Sr}/^{86}\text{Sr}_i$ and decreasing Nb and Ta. These gradients appear to gradually intersect the position of the convergence zone that lay seaward of the reconstructed MBL-Zealandia continental margin in pre-Cenozoic time. The pattern of anomalies suggests that the paleo-convergence zone lies offshore of western MBL and onshore in eastern MBL. This led to our interpretation that the Ba anomaly, and the Nd and Sr isotopic anomalies, are the product of residual subduction components in the mantle source.

5. The high-Ba anomalies, uncoupling of Ba from other LILE, and the erratic nature of the anomalies are most easily explained by some combination of hydrothermal activity, removal and uncoupling of Ba from the subduction mélange during slab de-watering, and

subsequent hydrous fluid transfer. Subsequent rise of hydrous fluids resulted in erratic, and often independent, Ba enrichment of mantle source rocks.

6. The low $^{143}\text{Nd}/^{144}\text{Nd}$ of Mt. Takahe (and one Mt. Siple) basaltic rocks are best explained by partial melting of slab mélange sediments in the mantle source that were richer in EM components than the sources of other MBL volcanoes.

7. Although Mt. Takahe, Mt. Murphy, Mt. Siple volcanoes, and the Jones Mountains, all display relict subduction signatures that seem related to the same convergent margin, the predominantly Pb-related anomalies in the Jones Mountains (high Pb/Ce, low $^{206}\text{Pb}/^{204}\text{Pb}$) are conspicuously different from the high-Ba and low $^{143}\text{Nd}/^{144}\text{Nd}$ anomalies that characterize the eastern, coastal MBL volcanoes. The geochemical discontinuity between the two regions coincides roughly with a tectonic boundary in the sea floor between the MBL sector, where subduction ceased ~85 Ma, and the Ellsworth Land sector, where subduction ended ~45 Ma. We speculate that a likely source of the discontinuity lies with differences in the composition of subducted sediments between the two regions.

8. We present new calculations that suggest a source melting depth of 80-90 km beneath Mt. Murphy. This roughly coincides with the base of the lithosphere, which was recently determined to be 80-100 km thick in MBL. We propose that the source of the anomalies resides with a fossil mélange diapir that rose from the Cretaceous subducting slab and became attached to the base of the lithosphere, where it remained during subsequent plate motions and Late Cenozoic remobilization, thereby preserving the subduction signature.

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875

876 **Figure captions.**

877 Figure 1. Index map, subglacial topography, and inferred structure (cross-section) of the Marie
878 Byrd Land dome and West Antarctic rift. The MBL dome is defined by contours on the
879 low relief West Antarctic erosion surface (WAES) (LeMasurier and Landis, 1996).
880 Structural relief on the dome from the inferred -500m base to crest (PT 2700) is ~ 3 km.
881 Ice-free, isostatically adjusted bedrock topography is simplified from Drewry (1983).
882 Abbreviations: AR Ames Range, B Mt. Bursey, C Mt. Cumming, CM Crary Mountains, CN

Coleman Nunatak, *E* Mt. Erebus, *ECR* Executive Committee Range, *F* Mt. Flint, *FR* Flood Range, *H* Mt. Hampton, *HB* Holmes Bluff, *HC* Hobbs Coast nunataks, *Hd* Hudson Mtns, *J* Jones Mtns, *M* Mt. Morning, *MBS* Marie Byrd Seamounts, *NVL* Northern Victoria Land, *P* Patton Bluff, *Pt* Mt. Petras, *RI* Ross Island, *SI* Shepard Island, *TAM* Transantarctic Mountains, *Tn* Toney Mountain, *VL* Victoria Land. Greenwich meridian is up, following the convention for small scale maps of Antarctica.

Figure 2. A plate tectonic and geologic reconstruction showing inferred spatial and geologic relationships between West Antarctica and Zealandia in pre-85 Ma time, modified from Cooper et al., (1982). Positions of the convergent plate margin and Pacific-Phoenix Ridge at ca. 105 Ma are adapted from Lawver and Gahagan (1994). The continuity of the Cretaceous plutonic belt and the Torlesse belt shows the overall geologic integrity of the reconstruction. West Antarctic erosion surface (WAES) data are from LeMasurier and Landis (1996). Abbreviations. *AI* Alexander Island, *Ak* Auckland city, *CI* Campbell Island, *C* Canterbury, *Ch* Chatham Islands, *F* Fiordland, *HC* Hobbs Coast nunataks, *J* Jones Mountains, *O* Otago, *P* Pacific-Phoenix Ridge, *STR* South Tasman Rise, *Tas* Tasmania, *TI* Thurston Island.

Figure 3. Geologic sketch map of Mt. Murphy, modified from LeMasurier and Thomson (1990, p.165) and Smellie (2001), showing numbered sample localities and $^{40}\text{Ar}/^{39}\text{Ar}$ ages from Wilch and McIntosh (2002). Sample numbers shown in Table 1 include year collected, numbered sample locality, and letters to distinguish between multiple samples collected

from the same locality, e.g. for sample number 85-32B: 85=year collected, 32=sample locality shown on this figure, B=one of several samples collected at this locality.

Figure 4. Geologic sketch map of Mt. Takahe (LeMasurier and Thomson, 1990, p. 170) showing numbered sample localities, and isotopic ages from Wilch et al., (2000). Sample numbers shown in Table 1 include year collected, numbered sample locality, and letters to distinguish multiple samples from the same locality, as in Fig. 3.

Figure 5. (A) IUGS classification of Mt. Takahe – Mt. Murphy – Mt. Siple basaltic rocks based on the total alkalis – silica (TAS) diagram (Le Bas et al., 1986). (B) TAS diagram including basaltic rocks from panel (A), plus basaltic samples from 15 other MBL volcanic centers (the “comparative suite”). **Red circles**, Mt. Takahe; **blue squares**, Mt. Murphy basal succession; **red squares**, Mt. Murphy parasitic cones; **blue triangle**, Mt. Siple basal succession; **red triangle**, Mt. Siple parasitic cone; **“X”**, Comparative suite: Ames/Flood Ranges, 4 volcanoes (LeMasurier et al., 2011); Hobbs Coast, 4 nunataks (Hart et al., 1997); Executive Committee Range, 3 volcanoes (LeMasurier et al., 2003); Mt. Flint, Mt. Petras, Shepard Island (LeMasurier and Thomson, 1990); Crary Mountains (Panter et al., 2000). See Fig. 1 for all locations. **M-K (1964)** is the boundary between tholeiitic and alkali basalts of Macdonald and Katsura (1964). Major elements recalculated to 100%, volatile free, following LeBas et al., (1986).

Figure 6. Major element variation diagrams for Mt. Takahe, Mt. Murphy, Mt. Siple, plus comparative suite. **a.** MgO vs Na₂O **b.** MgO vs Ni **c.** MgO vs CaO **d.** CaO/Na₂O vs V (Hart et al., 1997).

Figure 7. Multi-element plot of two **Mt. Takahe** and three **Mt. Murphy** basaltic rocks normalized to primordial mantle (McDonough and Sun, 1995). Included for comparison are basanites from the Ames Range (LeMasurier et al., 2011) and Jones **Mountains** (Hart et al., 1995).

Figure 8. Selected trace element plots of **Mt. Takahe-Mt. Murphy-Mt. Siple** basaltic rocks, compared with the comparative suite plus Jones **Mountains** (Hart et al. 1995). Vertical dashed lines on plots A, B, D, and F, roughly separate Hobbs Coast and Ames/Flood Range basalts in the west (see Fig 1) from **Mt. Takahe**, **Mt. Murphy**, and **the Jones Mountains** basalts in the east, to help illustrate the west to east Nb gradient. Clusters of most samples from **Mt. Murphy**, **Mt. Takahe**, Hobbs Coast, and AR/FR are encircled, where it can be done without adding clutter, to help visualize patterns described in the text. **(A)** Nb versus Ba. **(B)** Nb versus K. **(C)** K versus Ba/La **(D)** Ba/Nb versus K/Nb. HIMU and EM1 average end member points, here and in **Fig. 8E**, are from Hart and Gaetani (2006). The shaded area labeled AP is the range for Antarctic Peninsula basalts. AP and Hudson **Mountains** range are from Hole and LeMasurier (1994). **(E)** Ba/La versus Ba/Nb. **(F)** Nb versus Ta. Only ICPMS data are used for diagrams that involve La. The encircled samples are **Mt. Murphy** (M, in blue), **Mt. Takahe** (T, in red), Hobbs Coast (HC, in black), Ames and Flood Ranges (AR/FR, dotted line).

Figure 9. **a.** Pb/Ce versus Ba/Nb. **b.** Nb versus Th/Pb. **c.** Nb versus Nb/Y. Only ICPMS data are used for diagrams that involve Ce, Pb, and Th. OIB end members EM1, and HIMU are from Hart and Gaetani (2006). GLOSS is from Plank and Langmuir (1998). Encircled

samples are Mt. Murphy (M, in blue), Mt. Takahe (T, in red), and Jones Mountains (J, in green).

Figure 10. **a.** $^{143}\text{Nd}/^{144}\text{Nd}$ versus $(^{87}\text{Sr}/^{86}\text{Sr})_i$ isotopic ratios for Mt. Takahe, Mt. Murphy and Mt. Siple. The field “MBL basaltic rocks,” includes all available data from elsewhere in the province, from Futa and LeMasurier (1983); Hole and LeMasurier (1994); Hart et al., (1997); Panter et al., (2000). Jones Mountains points are from Hart et al., (1995). Fields for mid-ocean ridge basalts (MORB) and mantle components are from Zindler and Hart (1986): EM1 and EM2 enriched mantle of type 1 and 2 respectively, HIMU mantle with high U/Pb ratio. HIMU, EM1 and EM2 end member points are from Hart and Gaetani (2006). BSE is bulk silicate earth. Field for potential lower crustal contaminants is from Futa and LeMasurier, (1983), and Wysoczanski et al., (1995); the field for MBL upper crustal granitoids and gneisses is from Pankhurst et al., (1998); data for peridotite xenoliths from West Antarctica are from Martin et al., (2013), and for pyroxenite xenoliths from Wysoczanski et al., (1995); Perinelli et al., (2011); Martin et al., (2013). Error bars are 2σ , given only where they exceed the size of the symbol in the plot.

b. La/Nb versus $^{143}\text{Nd}/^{144}\text{Nd}$, modified from Hole and LeMasurier (1994), and showing the range of basalt compositions in West Antarctica, compared with western U.S. (Leat et al., 1988; Fitton et al., 1988).

c. $^{143}\text{Nd}/^{144}\text{Nd}$ versus $(^{87}\text{Sr}/^{86}\text{Sr})_i$ for Mt. Takahe, Mt. Murphy, and Mt. Siple basaltic rocks plus those from the comparative suite. Locations for all samples can be found in Fig. 1. Encircled samples as in Fig. 8.

Figure 11. $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$, and $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$, for Mt. Takahe, Mt. Murphy and Mt. Siple. Also shown are available Pb data for MBL lower crustal rocks (Wysoczanski et al., 1995), MBL granitoids and gneisses (Mukasa and Dalziel, 2000), and basaltic rocks from elsewhere in the MBL province (Hart et al., 1997; Panter et al., 2000) plus Jones Mountains (Hart et al., 1995), and peridotite and pyroxenite xenoliths from West Antarctica (Wysoczanski et al., 1995; Martin et al., 2013). Fields for oceanic basalts and mantle components are from Zindler and Hart (1986). EM1 and EM2 enriched mantle of type 1 and 2, respectively, HIMU mantle with high U/Pb ratio. MORB mid-ocean ridge basalts. HIMU, EM1, and EM2 end member points are from Hart and Gaetani (2006). NHRL is the Northern Hemisphere Reference Line of Hart (1984). Error bars are 2σ , given only where they exceed the size of the symbol in the plot.

Figure 12. Multi-element plot of high-Ba (85-32B, 67-66B) and low-Ba (67-62; 85-19) basaltic rocks from Mt. Murphy and Mt. Takahe, respectively. Also included are Ames Range sample AR60E, GLOSS (Plank and Langmuir, 1998) and Avg EM1 (Hart and Gaetani, 2006), all normalized to primitive mantle (McDonough and Sun, 1995).

KEY

- Early to middle Cretaceous Plutonism (mainly granitic)
- Remnants of the latest Cretaceous West Antarctic/ Waipounamu erosion surface
- Late Paleozoic-Jurassic quartzo-feldspathic graywackes (Torlesse)

111° W

110° W

MOUNT MURPHY

LEGEND

- Takahe
- Murphy basal succession
- Murphy parasitic cones
- x Comparative suite
- B Basalt
- BA Basaltic andesite
- Bn Basanite
- H Hawaiite
- M Mugearite
- Pb Picro basalt
- Pt Phonotephrite
- Tp Tephrite

A

 $^{143}\text{Nd}/^{144}\text{Nd}$

Table 1: LA-ICP-MS Settings

RF Power (W)	1200
Ar Nebulizer flow (L/min)	0.6-0.8
Ar Auxiliary gas (L/min)	0.7
Ar cooling gas (L/min)	15
He Carrier Gas (L/min)	0.3-0.7
ThO ⁺ /Th ⁺ (%)	<2
Laser-repetition rate (Hz)	8
Spot size (μm)	110
Energy Density (J/cm ²)	Matrix -Dependent (4.5-11)

Table 2: Average mineral phase major element concentrations from laser-ablation ICP-MS

Sample ID	Modal Reconstruction (%)	SiO ₂ wt%	TiO ₂ wt%	Al ₂ O ₃ wt%	Cr ₂ O ₃ wt%	FeO(t) wt%	MnO wt%	NiO wt%	MgO wt%	CaO wt%	Na ₂ O wt%	H ₂ O (Assumed) wt%	Total	MgO/SiO ₂
Clinopyroxene [@]	6	53.93	0.29	2.04	0.70	2.23	0.08	0.00	16.92	24.18	0.26		100.63	0.31
St. Dev.														
Orthopyroxene (n=1)	15	56.93		3.87	1.53	6.58	0.36	0.00	29.49	1.24			100.00	0.52
St. Dev.		0.26		0.12	0.11	0.41	0.13	0.00	0.22	0.16				
Serpentine (n=12)	75	42.51	0.02	0.57		3.81	0.06	0.26	39.73	0.03		13.00	100.38	0.94
St. Dev.		0.92	0.08	0.51		1.18	0.05	0.05	0.89	0.02				
Fe-Oxide [Ⓕ] (n=1)	2	20.89	0.03	1.20	1.77	60.93	0.08	0.13	14.94	0.02			100.00	
St. Dev.														
Plagioclase (n=2)	2	43.57	0.05	23.98		0.16	0.03	0.00	0.24	31.92	0.05		100.00	0.01
St. Dev.		2.56	0.00	1.24		0.09	0.03	0.00	0.18	1.04	0.01			
Reconstructed Bulk Rock		43.80	0.05	1.78	0.29	5.58	0.13	0.21	35.53	1.48	0.01	9.93	98.81	0.81
Clinopyroxene [@]		52.54	0.28	3.07	1.09	2.53	0.08		16.00	23.10	0.48			
Clinopyroxene (n=4)	7	51.11	0.15	5.02		3.93	0.10	0.08	20.54	19.08			100.00	0.40
St. Dev.		0.66	0.01	0.93		0.76	0.01	0.03	4.38	4.00				
Orthopyroxene (n=4)	25	55.27	0.07	4.74		6.49	0.14	0.09	31.41	1.78			100.00	0.57
St. Dev.		1.23	0.01	0.37		0.08	0.00	0.00	0.80	0.22				
Serpentine (n=6)	65	42.87	0.01	0.39		7.54	0.08	0.33	37.59	0.04		13.00	100.38	0.89
St. Dev.		1.21	0.02	0.14		1.41	0.03	0.04	1.09	0.01				
Fe-Oxide (n=2)	1	16.99	0.00	0.15		68.18	0.58	0.13	13.93	0.03	0.00		100.00	
St. Dev.		3.95	0.00	0.02		5.41	0.04	0.02	1.46	0.00	0.00			
Plagioclase* (n=2)	1	43.57	0.05	23.98		0.16	0.03	0.00	0.24	31.92	0.05		100.00	0.01

	St. Dev.		2.56	0.00	1.24	0.00	0.09	0.03	0.00	0.18	1.04	0.01			
	Reconstructed Bulk Rock		45.65	0.04	1.94	0.00	7.62	0.10	0.24	33.95	2.20	0.00	8.51	100.25	0.74
VAN 7-78-41	Clinopyroxene [@]		52.87	0.22	4.22	1.08	2.62	0.13		15.45	23.09	0.33			
	Clinopyroxene (n=2)	1	49.40	0.19	2.53		2.84	0.10	0.04	17.06	27.67			99.84	0.35
	St. Dev.		0.05	0.03	0.50		0.03	0.01	0.00	0.59	0.00				
	Orthopyroxene (n=3)	19	54.69	0.08	2.20		6.81	0.16	0.08	31.03	1.50			96.56	0.57
	St. Dev.		2.82	0.00	0.59		0.64	0.02	0.01	2.24	0.52				
	Serpentine (n=6)	77	42.57	0.00	0.50		2.72	0.06	0.35	40.76	0.03		13.00	100.76	0.91
	St. Dev.		2.73	0.09	0.97		2.34	0.09	0.13	3.85	0.33				
	Fe-Oxide (n=1)	2	20.89	0.03	1.20	1.77	60.93	0.08	0.13	14.94	0.02			100.00	
	St. Dev.														
	Plagioclase (n=2)	1	40.00	0.00	27.48		0.78	0.13	0.01	2.70	28.89			100.00	0.07
	St. Dev.		1.03	0.00	4.10		0.55	0.10	0.01	0.62	4.36				
	Reconstructed Bulk Rock		43.72	0.07	1.99	0.63	6.04	0.13	0.19	35.91	0.98	0.02	10.39	100.06	0.82
VAN 7-85-42	Clinopyroxene [§] (n=5)	5	51.65	0.22	6.28	1.08	3.27	0.10		17.06	21.09	0.36		101.11	0.33
	St. Dev.		0.38	0.03	0.40	0.11	0.14	0.01		0.35	0.57	0.01			
	Orthopyroxene [§] (n=3)	25	55.08	0.08	5.20	0.72	6.23	0.12		31.50	2.03	0.03		100.98	0.57
	St. Dev.		0.18	0.00	0.10	0.01	0.09	0.02		0.33	0.27	0.01			
	Serpentine (n=7)	69	41.48	0.01	0.40		7.34	0.07	0.32	37.33	0.05		13.00	100.41	0.86
	St. Dev.		2.64	0.04	1.32		4.35	0.06	0.08	2.48	0.20				
	Fe-Oxide [#]	1	16.37	0.00	0.47		65.82	1.68	0.36	15.24	0.07			100.00	
	St. Dev.														
	Reconstructed Bulk Rock		43.69	0.05	2.13	0.16	8.62	0.12	0.20	33.82	1.48	0.02	10.23	100.53	0.77

VAN 7-85-43	Clinopyroxene [§] (n=4)	8	51.79	0.17	6.33	1.13	2.97	0.11		16.68	21.03	0.40		100.61	0.32
	St. Dev.														
	Orthopyroxene (n=1)	25	54.86		7.20	1.22	6.13	0.14	0.16	28.34	1.92			99.97	0.52
	Serpentine (n=11)	66	41.13	0.00	0.43		7.22	0.06	0.33	37.80	0.03		13.00	100.33	0.92
	St. Dev.		0.54	0.00	0.13	0.00	0.83	0.01	0.02	0.41	0.00				
	Fe-Oxide (n=1)	1	16.37	0.00	0.47		65.82	1.68	0.36	15.24	0.07			100.00	
	Reconstructed Bulk Rock		45.01	0.02	2.31	0.37	7.34	0.10	0.27	33.85	2.07	0.03	8.90	100.27	0.75
VAN 7-85-47	Clinopyroxene [§]		51.54	0.22	6.31	1.07	3.21	0.10		17.08	20.88	0.37			
	Clinopyroxene (n=6)	5	51.17	0.16	5.76		3.54	0.11	0.05	18.25	20.64			99.69	0.36
	St. Dev.		1.15	0.01	0.42		2.37	0.00	0.00	0.92	1.45				
	Orthopyroxene (n=1)	16	56.09	0.07	4.69		6.18	0.14	0.09	30.93	1.80			100.00	0.55
	Serpentine (n=6)	78	42.46	0.01	1.16		7.99	0.09	0.27	34.76	0.26		13.00	100.38	0.84
	St. Dev.		2.21	0.01	1.16		2.37	0.02	0.06	1.49	0.34				
	Fe-Oxide [#]	1	16.37	0.00	0.47		65.82	1.68	0.36	15.24	0.07			100.00	
	Reconstructed Bulk Rock		44.61	0.03	1.69	0.00	8.33	0.11	0.24	33.45	1.63	0.00	10.19	100.28	0.75

"n" equals the number of spots determined.

⌘ Assumed magnetite from VAN 7-78-41.

Assumed magnetite from VAN 7-85-43.

* Assumed plagioclase from VAN 7-78-36H.

@ Major element data from Mallick et al., (2014).

§ Major element data from Warren et al., (2009).

Table 3: Average mineral phase trace element concentrations from laser-ablation ICP-MS

Sample ID		Ba	Th	U	La	Ce	Pb	Pr	Nd	Sr	Zr	Hf	Sm	Ti	Gd	Dy	Y	Er	Yb	Lu
		(ng/g)	(ng/g)	(ng/g)	(ng/g)	(ng/g)	(ng/g)	(ng/g)	(ng/g)	(ug/g)	(ng/g)	(ng/g)	(ng/g)	(ug/g)	(ng/g)	(ng/g)	(ng/g)	(ng/g)	(ng/g)	(ng/g)
VAN 7-78-36H	Bulk Rock	580.6	8.12	425.4	101.8	271.2	29.87	42.89	226.6	34.48	1823.6	59.65	95.60	320.5	169.4	248.6	1602	185.7	201.9	32.35
	Clinopyroxene [@]	0.00	7.27	52.89	310.0	1490	59.70	0.00	2200	1.780	14810	480	1020	2093	1940	3090	17180	1930	1710	220.0
	Orthopyroxene (n=2)	0.95	8.02	3.11	3.98	25.10	10.73	5.50	38.63	0.014	2805.5	79.63	22.54	869.6	73.38	230.9	1858	292.6	455.3	79.85
	St. Dev.	4.38	4.47	1.10	3.02	24.51	0.96	5.45	39.50	0.002	2445.9	59.40	17.82	29.09	62.60	137.1	960.3	134.8	136.2	19.55
	Serpentine (n=12)	241.2	3.42	878.7	27.17	64.84	32.63	12.01	68.94	1.491	625.13	16.60	29.24	119.5	51.80	84.44	580.0	73.93	98.85	18.31
	St. Dev.	207.9	1.39	426.9	13.00	52.09	12.76	6.34	41.38	0.370	469	17.69	18.22	63.70	32.6	48.9	285	37.0	32.2	5.3
	Fe-Oxide [Ⓐ] (n=2)	136.8	10.42	208.9	21.64	67.50	38.41	13.07	54.42	0.997	589.80	12.19	21.09	4349	20.70	33.46	162.7	22.13	22.73	5.85
	St. Dev.	544.7	4.82	173.0	9.78	3.78	230.7	7.13	2.39	0.136	44.90	1.65	2.68	0.05	2.78	7.98	94.26	8.07	4.71	1.78
	Plagioclase (n=2)	3423	1.53	13.48	1295	2290	387.1	231.99	785.4	73.79	0.44	2.12	100.5	323.2	79.60	32.98	158.4	12.54	7.31	2.51
	St. Dev.	1179	0.55	7.08	268.7	244.2	11.65	29.81	114.5	19.17	0.00	1.42	9.71	28.38	0.88	1.58	4.81	4.62	1.18	1.84
	Reconstructed Bulk Rock	184.7	4.02	659.9	40.13	143.6	26.61	30.50	191.2	1.239	1788.9	53.59	87.44	350.8	167.8	285.8	1757	216.7	246.5	39.21
VAN 7-78-39	Bulk Rock	407.0	7.27	52.89	51.53	110.4	59.70	16.83	77.49	2.500	729.78	29.73	46.02	262.9	98.96	225.7	1548	194.0	224.3	37.08
	Clinopyroxene (n=4)	64.46	2.88	54.40	4.98	11.35	15.77	14.68	247.1	0.365	495.27	117.9	389.2	945.0	1034	1838	10788	1299	1221	177.9
	St. Dev.	64.30	4.26	75.33	7.71	6.95	18.69	6.56	10.96	0.221	45.95	12.95	7.97	44.59	60.68	32.65	287.9	59.20	21.35	3.93
	Orthopyroxene (n=4)	20.47	0.61	15.84	5.40	3.25	28.19	0.80	9.23	0.148	115.68	23.47	20.37	440.7	68.53	223.9	1795	261.4	385.5	69.56
	St. Dev.	20.68	0.78	20.61	8.41	4.32	15.03	1.69	5.37	0.206	10.02	3.23	5.86	36.07	17.35	18.95	141.3	19.12	28.61	3.00
	Serpentine (n=6)	454.9	1.24	1401	2.79	2.14	59.17	1.38	6.99	2.183	31.19	2.65	9.64	48.89	31.66	63.80	491.5	63.67	81.87	15.96
	St. Dev.	109.6	0.99	1603	3.38	1.89	44.85	1.43	3.38	1.1	14	1.33	6.16	9.41	29.3	60.0	523	68.4	85.0	15.5
	Fe-Oxide (n=2)	1312	4.80	424.7	9.82	15.68	95.99	5.64	17.01	3.630	89.01	7.48	23.34	9.66	48.05	75.42	416.9	61.31	54.83	12.57
	St. Dev.	544.7	4.82	173.0	9.78	3.78	230.7	7.13	2.39	0.136	44.90	1.65	2.68	0.05	2.78	7.98	94.26	8.07	4.71	1.78
	Plagioclase* (n=2)	3423	1.53	13.48	1295	2290	387.2	231.99	785.4	73.79	0.44	2.12	100.5	323.2	79.60	32.98	158.4	12.54	7.31	2.51
	St. Dev.	1179	0.55	7.08	268.7	244.2	11.65	29.81	114.5	19.18	0.00	1.42	9.71	28.38	0.88	1.58	4.81	4.62	1.18	1.84
	Reconstructed Bulk Rock	366.2	1.38	929.9	7.42	7.42	48.43	2.76	26.75	2.085	87.19	16.54	40.85	212.7	115.5	235.4	1577	204.3	241.4	41.28
VAN 7-78-41	Bulk Rock	755.3	4.62	651.7	53.32	81.78	40.65	12.61	53.12	3.110	493.95	19.28	34.35	185.7	71.85	173.1	1271	150.1	169.9	29.79
	Clinopyroxene (n=2)	195.2	39.60	20.06	385.7	1214	18.75	213.17	1167	0.786	6825.3	274.3	764.5	1140	1664	2862	16529	1927	1736	239.0

	St. Dev.	87.26	1.32	0.73	36.40	8.38	0.83	29.45	239.2	0.026	2309.4	56.57	192.7	183.4	330.9	441.7	2737	292.5	282.3	15.61
	Orthopyroxene (n=3)	35.59	24.32	16.34	23.48	50.11	23.51	22.28	46.79	0.124	970.89	55.55	44.79	481.3	110.7	293.4	2395	359.4	542.5	104.9
	St. Dev.	5.82	13.10	9.66	13.05	30.55	12.91	9.25	25.80	0.041	368.69	24.60	19.33	25.30	43.71	61.10	252.0	46.25	50.01	15.15
	Serpentine (n=5)	359.9	7.58	160.7	56.12	59.08	30.33	20.80	53.33	1.205	147.19	9.42	21.19	27.10	34.87	51.93	322.3	48.09	64.84	15.68
	St. Dev.	23.30	3.07	120.4	13.50	38.87	12.69	4.81	18.99	0.2	112	8.69	12.08	14.04	29.4	44.8	278	36.1	30.1	7.9
	Fe-Oxide (n=1)	188.8	12.88	27.19	17.23	37.18	26.82	11.84	8.13	0.585	31.00	3.13	10.51	193.2	9.48	12.32	43.64	13.74	17.50	7.62
	Plagioclase (n=2)	4403	15.35	18.76	376.9	418.9	40.44	50.73	134.9	55.20	11.15	10.01	31.33	24.31	41.96	36.89	162.8	21.04	17.29	8.86
	St. Dev.	2427	18.14	17.02	81.28	96.09	47.53	12.55	32.57	30.01	9.37	5.81	23.98	8.39	36.91	3.23	22.16	1.61	12.47	0.39
VAN 7-85-42	Reconstructed Bulk Rock	333.9	11.30	127.7	56.31	74.28	28.94	23.48	65.25	1.524	377.99	21.14	34.35	129.2	68.07	129.9	898.3	128.1	173.3	34.92
	Bulk Rock	698.7	1.18	745.6	3.39	4.16	3.56	1.21	15.19	143.3	149.77	19.87	26.31	186.6	64.29	160.7	1240	144.1	167.9	29.40
	Clinopyroxene [§] (n=3)	7.63	0.03	6.65	0.20	5.11	13.59	8.85	201.1	0.173	410.01	87.65	344.5	962.8	877.2	1633	9416	1151	1090	156.7
	St. Dev.	6.17	0.00	10.65	0.00	0.16	7.34	0.70	4.07	0.081	14.53	3.56	14.17	23.31	46.49	42.29	269.5	10.96	36.50	4.15
	Orthopyroxene [§] (n=3)	3.46	0.04	0.91	0.30	0.25	8.18	0.08	3.90	0.023	90.80	21.14	13.22	434.8	56.27	181.2	1460	205.5	324.9	58.33
	St. Dev.	3.80	0.05	0.89	0.00	0.18	3.52	0.10	0.74	0.044	7.31	2.10	2.10	9.09	4.89	8.93	64.41	8.57	14.30	3.69
	Serpentine (n=7)	108.5	0.09	1079	0.14	0.28	14.41	0.24	1.90	3.1	30	3.93	5.36	42.82	16.8	38.0	303	36.6	54.4	10.5
	St. Dev.	74.15	0.08	993.9	0.08	0.21	6.31	0.22	2.66	1.9	25	5.90	6.82	55.30	19.1	39.6	275	32.6	48.2	10.2
VAN 7-85-43	Fe-Oxide [#]	9186	2.45	495.1	363.7	352.8	421.7	111.18	429.0	14.89	157.19	1.77	133.3	5.93	166.6	216.9	908.2	141.6	130.6	21.97
	Reconstructed Bulk Rock	157.6	0.10	754.9	3.39	3.60	16.41	1.59	15.55	2.266	64.43	12.14	24.42	183.5	68.46	150.4	1025	132.1	171.3	29.40
	Bulk Rock	374.5	0.48	948.5	17.37	3.51	64.31	3.98	32.73	3.561	149.44	18.41	35.60	179.5	93.57	192.9	1425	163.1	181.7	30.87
	Clinopyroxene [§] (n=4)	2.00	0.00	2.67	0.16	5.69	12.26	10.21	223.2	0.160	356.14	82.04	354.8	847.1	844.4	1551	8896	1083	1059	151.2
	St. Dev.	0.00	0.03	2.96	0.09	0.76	5.71	0.72	10.11	0.008	70.73	5.30	14.70	17.17	23.66	34.92	139.1	17.96	30.58	3.25
	Orthopyroxene (n=3)	12.90	0.02	37.37	1.41	0.65	13.26	0.66	8.31	0.087	100.87	18.71	16.81	383.3	69.13	192.5	1470	214.1	319.0	57.44
	St. Dev.	21.81	0.03	62.95	1.91	0.44	2.30	1.06	4.29	0.149	41.08	3.46	2.87	38.48	18.78	29.26	177.3	21.57	25.97	4.49
	Serpentine (n=11)	107.8	0.23	1094	7.68	0.99	33.96	2.57	11.78	1.895	105.17	3.11	6.76	15.63	17.36	33.57	240.5	30.37	42.15	7.70
	St. Dev.	65.11	0.43	132.9	6.61	2.35	37.04	1.72	7.24	0.4	98	4.03	2.66	17.47	5.8	10.3	67	7.4	12.8	1.9
	Fe-Oxide (n=1)	9186	2.45	495.1	363.7	352.8	421.7	111.18	429.0	14.89	157.19	1.77	133.3	5.93	166.6	216.9	908.2	141.6	130.6	21.97

	Reconstructed Bulk Rock	446.5	0.91	752.3	8.51	4.29	10.94	2.41	28.49	3.835	101.73	13.65	33.53	154.9	89.09	169.9	1189	138.1	158.4	25.67
VAN 7-85-47	Bulk Rock	431.9	1.13	786.8	7.28	4.62	6.53	2.39	25.67	4.039	137.67	16.82	34.95	170.6	79.75	192.9	1447	162.2	177.9	30.93
	Clinopyroxene (n=6)	16.76	2.36	7.93	3.25	9.70	17.74	15.25	273.4	0.263	455.52	112.9	436.2	942.9	1092	1920	11342	1345	1277	183.2
	St. Dev.	16.26	1.33	2.94	1.91	4.12	22.92	2.67	26.18	0.070	31.60	15.60	22.20	65.98	84.28	91.80	614.0	76.09	101.3	12.71
	Orthopyroxene (n=1)	15.80	0.87	7.64	1.36	1.17	6.94	0.82	9.21	0.141	102.05	24.59	16.97	413.1	77.10	202.9	1559	239.0	334.7	54.79
	Serpentine (n=6)	225.7	2.14	1668	1.88	2.22	14.70	1.87	4.99	7.687	48.48	8.61	4.82	88.86	16.66	41.61	323.8	40.09	71.25	13.55
	St. Dev.	82.59	1.31	623.0	1.70	1.29	7.86	1.57	1.25	6.2	21	5.82	2.61	99.68	8.4	24.3	189	23.1	43.9	8.8
	Fe-Oxide [#]	9186	2.45	495.1	363.7	352.8	421.7	111.18	429.0	14.89	157.19	1.77	133.3	5.93	166.6	216.9	908.2	141.6	130.6	21.97
	Reconstructed Bulk Rock	233.1	1.95	1312	3.96	4.48	16.00	3.06	22.60	6.145	79.51	16.72	30.79	185.4	85.17	169.5	1116	142.5	178.1	29.23

"n" equals the number of spots determined.

⌘ Assumed magnetite from VAN 7-78-41.

Assumed magnetite from VAN 7-85-43.

* Assumed plagioclase from VAN 7-78-36H.

@ Major element data from Mallick et al., (2014).

§ Major element data from Warren et al., (2009).

Table 4: Peridotite leachates and residual rock trace element concentrations

Sample ID		Ba	Th	U	La	Ce	Pb	Pr	Nd	Sr	Zr	Hf	Sm	Ti	Gd	Dy	Y	Er	Yb	Lu
		(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)	(ug/g)
VAN 7-78-36H	H ₂ O Leach	0.008	0.000	0.006	0.001	0.002	0.001	0.000	0.001	2.296	0.006	0.000	0.001	1.149	0.001	0.002	0.008	0.001	0.001	0.000
	NaOAc Leach	0.248	0.000	0.215	0.017	0.033	0.001	0.005	0.029	17.549	0.000	0.000	0.012	0.039	0.021	0.030	0.164	0.022	0.022	0.003
	HH Leach	0.141	0.001	0.104	0.011	0.026	0.011	0.004	0.018	2.506	0.002	0.000	0.007	0.171	0.012	0.019	0.100	0.014	0.018	0.003
	6N HCl Leach	0.102	0.003	0.075	0.032	0.084	0.006	0.013	0.066	0.757	0.098	0.002	0.027	49.363	0.050	0.074	0.406	0.055	0.061	0.010
	Residual Rock	0.090	0.003	0.028	0.030	0.090	0.015	0.015	0.082	0.365	1.369	0.043	0.036	172.11	0.064	0.098	0.613	0.070	0.073	0.012
VAN 7-78-39	H ₂ O Leach	0.010	0.000	0.007	0.000	0.000	0.000	0.000	0.000	0.160	0.002	0.000	0.000	0.635	0.000	0.001	0.005	0.001	0.001	0.000
	NaOAc Leach	0.357	0.000	0.231	0.008	0.007	0.002	0.002	0.008	0.775	0.000	0.000	0.004	0.042	0.008	0.015	0.094	0.012	0.013	0.002
	HH Leach	0.232	0.000	0.106	0.006	0.007	0.010	0.001	0.006	0.340	0.001	0.000	0.003	0.074	0.007	0.011	0.070	0.009	0.013	0.002
	6N HCl Leach	0.060	0.001	0.136	0.013	0.020	0.007	0.003	0.014	0.340	0.026	0.001	0.008	35.400	0.020	0.041	0.248	0.036	0.041	0.007
	Residual Rock	0.138	0.003	0.005	0.025	0.056	0.018	0.008	0.041	0.196	0.527	0.022	0.023	146.02	0.056	0.108	0.719	0.083	0.093	0.016
VAN 7-78-41	H ₂ O Leach	0.005	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.137	0.002	0.000	0.000	0.758	0.000	0.000	0.003	0.000	0.001	0.000
	NaOAc Leach	0.164	0.000	0.025	0.006	0.007	0.001	0.001	0.006	0.433	0.001	0.000	0.003	0.035	0.005	0.010	0.067	0.010	0.012	0.002
	HH Leach	0.071	0.001	0.009	0.005	0.006	0.007	0.001	0.004	0.216	0.000	0.000	0.002	0.155	0.003	0.006	0.035	0.005	0.008	0.001
	6N HCl Leach	0.106	0.002	0.015	0.023	0.044	0.005	0.005	0.023	0.567	0.035	0.001	0.012	41.137	0.029	0.055	0.332	0.046	0.053	0.010
	Residual Rock	0.160	0.003	0.083	0.020	0.032	0.042	0.004	0.021	0.317	0.419	0.016	0.013	96.597	0.031	0.066	0.440	0.054	0.069	0.012
VAN 7-85-42	H ₂ O Leach	0.005	0.000	0.010	0.000	0.000	0.000	0.000	0.000	0.246	0.001	0.000	0.000	0.313	0.000	0.000	0.002	0.000	0.000	0.000
	NaOAc Leach	0.398	0.000	0.417	0.001	0.000	0.000	0.000	0.002	70.086	0.001	0.000	0.003	0.067	0.006	0.014	0.093	0.013	0.014	0.003
	HH Leach	0.110	0.000	0.193	0.003	0.001	0.005	0.001	0.003	0.523	0.002	0.000	0.002	0.186	0.004	0.006	0.044	0.006	0.008	0.001
	6N HCl Leach	0.118	0.000	0.068	0.001	0.001	0.001	0.001	0.008	0.277	0.013	0.001	0.011	43.816	0.028	0.056	0.325	0.045	0.046	0.007
	Residual Rock	0.117	0.001	0.015	0.006	0.003	0.015	0.001	0.007	0.356	0.112	0.010	0.013	99.961	0.035	0.080	0.547	0.063	0.082	0.014
VAN 7-85-43	H ₂ O Leach	0.004	0.000	0.007	0.000	0.000	0.000	0.000	0.000	0.248	0.001	0.000	0.000	0.484	0.000	0.000	0.003	0.000	0.001	0.000
	NaOAc Leach	0.190	0.000	0.490	0.005	0.000	0.000	0.001	0.005	1.360	0.000	0.000	0.003	0.045	0.009	0.016	0.116	0.015	0.016	0.003
	HH Leach	0.110	0.000	0.221	0.006	0.001	0.007	0.001	0.006	0.411	0.002	0.000	0.004	0.172	0.009	0.017	0.102	0.014	0.019	0.003
	6N HCl Leach	0.058	0.000	0.102	0.002	0.001	0.003	0.001	0.008	0.173	0.007	0.001	0.010	48.134	0.028	0.055	0.328	0.044	0.047	0.008
	Residual Rock	0.083	0.000	0.036	0.005	0.002	0.010	0.001	0.009	0.090	0.076	0.009	0.013	83.144	0.038	0.075	0.519	0.061	0.070	0.013
VAN7-85-47	H ₂ O Leach	0.011	0.000	0.010	0.000	0.000	0.000	0.000	0.000	5.049	0.000	0.000	0.000	0.788	0.000	0.001	0.004	0.001	0.001	0.000
	NaOAc Leach	0.180	0.000	0.430	0.002	0.000	0.002	0.000	0.002	1.512	0.001	0.000	0.001	0.035	0.003	0.007	0.056	0.006	0.007	0.002
	HH Leach	0.148	0.000	0.186	0.001	0.001	0.004	0.000	0.001	11.216	0.002	0.000	0.002	0.155	0.005	0.011	0.074	0.012	0.016	0.003
	6N HCl Leach	0.063	0.000	0.042	0.000	0.001	0.002	0.000	0.003	0.665	0.007	0.001	0.005	37.985	0.012	0.028	0.171	0.025	0.028	0.005
	Residual Rock	0.056	0.001	0.033	0.005	0.002	0.011	0.001	0.012	0.122	0.079	0.011	0.019	146.02	0.047	0.092	0.592	0.073	0.078	0.014

Table 5: $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd} Isotopes for Bulk Rock, primary CPX, Leachates and Residual rock.

		$^{87}\text{Sr}/^{86}\text{Sr}$	2σ	ϵ_{Nd}	2σ	% SW Nd contribution [@]
VAN 7-78-36H	Bulk Rock	0.708543	0.000005	7.26	0.10	
	NaOAc + HH Leach	0.709108	0.000004	7.78	0.06	
	6N HCl Leach	N.D.		5.17	1.84	3
	Residual rock	0.705788	0.000020	8.00	0.06	1
	CPX	0.702858	0.000019	7.92	0.37	14
VAN 7-78-39	Bulk Rock	0.707067	0.000007	6.85	0.08	0
	NaOAc + HH Leach	N.D.		3.15	0.23	
	6N HCl Leach	N.D.		5.05	0.08	4
	Residual rock	0.706889	0.000006	6.43	0.10	24
	CPX	0.702992	0.000007	7.67	0.18	14
VAN 7-78-41	Bulk Rock	0.708411	0.000007	5.46	0.08	6
	NaOAc + HH Leach	0.709039	0.000028	2.98	0.14	
	6N HCl Leach	N.D.		6.84	0.10	13
	Residual rock	0.706889	0.000006	5.56	0.10	26
	CPX	0.707895	0.000008	7.96	0.41	6
VAN 7-85-42	Bulk Rock	0.709175	0.000008	10.26	0.18	12
	NaOAc + HH Leach	0.709163	0.000003	6.42	0.25	
	6N HCl Leach	N.D.		11.17	0.24	12
	Residual rock	0.709463	0.000008	10.59	0.12	28
	CPX	0.702440	0.000017	13.28	0.27	9
VAN 7-85-43	Bulk Rock	0.709143	0.000008	5.70	0.27	11
	NaOAc + HH Leach	N.D.		-1.81	0.14	
	6N HCl Leach	N.D.		8.08	1.12	31
	Residual rock	0.709044	0.000010	9.54	1.35	61
	CPX	0.702902	0.000147	13.30	0.41	21
VAN 7-85-47	Bulk Rock	0.709199	0.000008	10.24	0.55	15
	NaOAc + HH Leach	0.709250	0.000009	-2.80	0.22	
	6N HCl Leach	N.D.		10.33	0.11	13
	Residual rock	0.709568	0.000007	11.12	0.14	65
	CPX	0.702308	0.000013	13.36	0.06	12

* From Table 1 Station 101 at 4400m depth in Stichel et al (2012).

@ Calculated from equation 3 using measured bulk rock, CPX and seawater ϵ_{Nd} .

Table 6: Water/Rock ratio bulk rock and sans CPX models

	Sr Conc (ppm)	Nd Conc (ppm)	Primary Magmatic $^{87}\text{Sr}/^{86}\text{Sr}^{\S}$	Primary Magmatic $\epsilon_{\text{Nd}}^{\S}$	Bulk Rock $^{87}\text{Sr}/^{86}\text{Sr}$	Bulk Rock ϵ_{Nd}	Closed BR W/R $^{\sim}$	Open BR W/R $^{\sim}$	Closed BR W/R $^{\#}$	Open BR W/R $^{\#}$	Nd addition W/R $^{\text{c}}$
SWIR Seawater*	8	2.00x10 ⁻⁶			0.70916	-9.70					
Van 7-78-36H BR	34.47	0.23	0.702858	7.92	0.708543	7.26	39.72	10.02	-	-	-
Van 7-78-36H BR without CPX	34.26	0.17				7.04	41.02	10.14	-	-	
Van 7-78-39 BR	2.50	0.08	0.702992	7.67	0.707067	6.85	0.61	0.34	-	-	-
Van 7-78-39 BR without CPX	2.45	0.06				6.59	0.63	0.34	-	-	
Van 7-78-41 BR	3.11	0.05	0.702800	5.56	0.708411	5.46	2.91	0.83	-	-	-
Van 7-78-41 BR without CPX	3.11	0.05				5.41	2.92	0.83	-	-	
Van 7-85-42 BR	143.35	0.02	0.702440	13.28	0.709175	10.26	-	-	1150.27	1071.06	927.52
Van 7-85-42 BR without CPX	143.33	0.01				4.99	-	-	1562.87	1239.12	
Van 7-85-43 BR	3.56	0.03	0.702902	13.30	0.709143	5.70	-	-	8085.29	6570.33	5024.71
Van 7-85-43 BR without CPX	3.54	0.02				-2.72	-	-	17829.42	9263.89	
Van 7-85-47 BR	4.04	0.03	0.702308	13.36	0.709199	10.24	-	-	2229.37	2071.21	1790.26
Van 7-85-47 BR without CPX	4.01	0.01				6.88	-	-	2680.59	2262.90	

$\S$ Derived from primary clinopyroxene for each sample.

* From Table 1 Station 101 at 4400m depth in Stichel et al (2012).

$\sim$ W/R determining from Sr isotopes using equations used from Delacour et al (2008).

$\#$ W/R determining from Nd isotopes using equations used from Delacour et al (2008).

c Addition from the determined Nd difference between the measured bulk rock and reconstructed primary peridotite.