The Aluminum Conundrum in Hawaiian Shield-Building Lavas: An Argument for a Deep, Garnet-Bearing, Mantle Source.

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Abstract

In contrast with other major elements (e.g. SiO₂, CaO, TiO₂), isotopic ratios (e.g. Sr, Nd, Pb), and some trace element ratios (e.g. Zr/Nb, Zr/Y, La/Nb), the Al₂O₃ content of Hawaiian shield lavas, at a given MgO content, is remarkably uniform within and between volcanoes. The overall range at 10 percent MgO is about 0.8% from Loihi (12.1 %) to Koolau (12.9 %). There is substantial overlap between the “Loa” and “Kea” trend volcanoes, but the regression data for “Loa” trend volcanoes are slightly higher (0.2%) in Al₂O₃ than “Kea” trend volcanoes. Post-shield lavas, from both “Kea” and “Loa” trends are higher in Al₂O₃ at a given MgO content than corresponding shield lavas. These observations are completely at odds with what we think we know about melt production in a heterogeneous, thermally zoned, mantle plume. Peridotite melting models and experiments, at varying pressures and temperatures, result in melts with significantly different Al₂O₃ contents at a given MgO content. The almost constant Al₂O₃ at a given MgO content is analogous to the almost constant normalized Yb abundances in Hawaiian tholeiites. Both require melting in the presence of a garnet residue. The aluminum conundrum can be explained if Hawaiian shield lavas result from melting garnet peridotite at depths corresponding to around 120 km. However, this explanation is at odds with the variable, but relatively high SiO₂ contents of most of these lavas, which imply melt production at relatively shallow depths corresponding to less than 90 km. This paradox can be resolved if low SiO₂ parental magmas derived from garnet peridotite react with a depleted harzburgite residue from prior melting as the magma ascends through the upper regions of the plume.

1. Introduction

The compositions of basaltic magmas erupted from Hawaiian volcanoes are widely interpreted to be a consequence of melting and magma production as individual volcanoes are transported by the northwest-moving Pacific plate over a stationary, thermally-zoned and compositionally-heterogeneous mantle plume [e.g. *Watson and Mackenzie*, 1991; *Frey and Rhodes*, 1993; *Ribe and Christensen*, 1999; *Farnetani and Hofmann*, 2010]. The most compelling evidence for this inference is found in the compositional differences between shield-stage and post- and pre-shield stage volcanism [*Stearns and Macdonald*, 1946; *Macdonald et al.*, 1983; *Moore and Clague*,

1992]. Shield- stage lavas are tholeiitic, whereas post- and pre-shield stage lavas are often alkalic, or characterized by transitional tholeiites with lower SiO₂ and higher alkali and incompatible element content than shield-stage lavas. These differences are thought to be indicative of lower degrees of magma production at the cooler margins of the plume, an inference supported not only by the presence of alkalic and low SiO₂ lavas with high incompatible element abundances, but also by a greater preponderance of evolved lavas in the post-shield stage [e.g. *Frey et al.*, 1991; *Feigenson et al.*, 1983], accompanied by almost an order of magnitude reduction in lava accumulation rates [*Sharpe and Renne*, 2005].

Therefore, as a volcano traverses the thermally-zoned and compositionally-heterogeneous mantle plume, we should expect to find significant differences in: melt production between volcanoes; melt production over time at individual volcanoes; the depth of melt segregation, within and between volcanoes; and the proportion of source components, within and between volcanoes. To a large extent these expectations are realized, given the difficulties in sampling the ~100 ka magmatic histories of young Hawaiian volcanoes [e.g. *DePaolo and Stolper*, 1996; *Quane et al.*, 2000; *Stolper et al.*, 2009; *Rhodes*, 2015]. There are significant differences between individual volcanoes and, sometimes, during the growth of an individual volcano, in major elements, trace elements and isotopic ratios [e.g. *Wright*, 1971; *Staudigal et al.*, 1984; *Garcia et al.*, 1989; *Rhodes et al.*, 1989; *Lassiter et al.*, 1996; *Frey et al.*, 2001; *Frey and Rhodes*, 1993; *Rhodes and Vollinger*, 2004; *Stolper et al.*, 2004; *Abouchami et al.*, 2005; *Weis et al.*, 2011; *Jackson et al.*, 2012].

The differences in major element composition of lavas erupted during the shield-building stage of individual Hawaiian volcanoes has long been recognized [e.g. *Powers*, 1955; *Wright*, 1971;

Rhodes et al., 1989; *Garcia et al.*, 1989; *Moore and Clague*, 1992; *Frey and Rhodes*, 1993]. In particular, volcanoes along the westerly “Loa” and easterly “Kea” trends of *Jackson et al.* (1972) differ significantly from one another. For example, at a given MgO content, “Kea” trend lavas tend to be lower in SiO₂, and higher in CaO, TiO₂, Na₂O and K₂O than “Loa” trend lavas. They also tend to be higher in highly incompatible trace element abundances. See figures 3 to 6 in *Jackson et al.* [2012] for a detailed summary of major element and isotopic differences between Hawaiian volcanoes and the “Loa” and “Kea” trends. Lower abundances of MgO-normalized TiO₂ and other incompatible major and trace elements are generally attributed to a greater extent of melting needed to produce the “Loa” trend magmas compared to the Kea trend magmas [e.g. *Hofmann and Farnetani*, 2013]. In contrast, normalized SiO₂ is pressure-dependant, relating to the depth of melt segregation from a peridotite source [e.g. *Carmichael et al.*, 1970; *Albarède*, 1992; *Lee et al.*, 2009], and also to the extent of peridotite melting [e.g. *Kushiro*, 1996; *Walter*, 1998; *Longhi*, 2002]. On the other hand, the apparent correlation of normalized SiO₂ abundances with isotopic ratios has been used as an argument for varying amounts of a subducted oceanic crustal eclogite/pyroxenite component, mixed with a peridotite component, in the plume source of Hawaiian volcanoes [e.g. *Hauri*, 1996; *Takahashi and Nakajima*, 2002; *Sobolev et al.*, 2005; *Jackson et al.*, 2012]. *Herzberg* [2006] takes this even further, arguing that only high CaO - low SiO₂ lavas from Loihi and Mauna Kea are derived from peridotite melting, all others are produced by melting of an olivine-free eclogite/pyroxenite source.

What is surprising is that, despite the well-documented variations in MgO normalized SiO₂, CaO and TiO₂, the MgO normalized abundances of Al₂O₃ are remarkably uniform in Hawaiian shield-building lavas, with very little difference between individual volcanoes, or between “Kea” and

“Loa” trend volcanoes. This presents a conundrum, because examination of the results of peridotite melting experiments leads one to expect that the Al_2O_3 content of a melt will be sensitive to the extent of melting as well as to the depth at which the melt segregates from its residue [e.g. *Kushiro*, 1996; *Walter*, 1998; *Longhi*, 2002; *Salters et al.*, 2002]. Additionally, assuming much of the heterogeneity inferred for the Hawaiian plume results from mixtures of peridotite and subducted oceanic crustal components, one might anticipate considerable variation in Al_2O_3 content in Hawaiian shield-building lavas. Indeed, *Sobolev et al.*, [2005, Fig. 4] use the Al_2O_3 content of inferred primary magmas to estimate the proportions of melt derived from a pyroxenitic component in the Hawaiian plume. In this paper, I will document the extraordinary uniformity of Al_2O_3 among Hawaiian shield-building lavas from volcanoes along both the “Loa” and “Kea” trends, arguing that the only plausible reason for this uniformity is the presence of substantial garnet in the source peridotite. This implies that these magmas were derived by melting of garnet peridotite in the Hawaiian plume at depths greater than 90 km and most probably closer to about 120 km.

2. Analytical Methods

About 200 - 600 g of sample were crushed by hand in a high-carbon steel percussion mortar. The resulting gravel-sized fragments were washed in tap water, followed by washing in an ultrasonic bath with de-ionized water. About 100 - 200 g was further crushed to about 200 mesh size in a Spex tungsten carbide (WC) shatterbox. Aliquants of this powder were analyzed by X-ray fluorescence analysis (XRF). The major elements were measured on a fused La-bearing lithium borate glass disc using modifications of the method of *Norrish and Hutton* [1969], the principle difference being that the samples are first ignited at 1000 - 1040 °C for several hours in order to

oxidize iron to Fe^{3+} and remove volatiles. They are then fused in an electric furnace at $\sim 1020^\circ\text{C}$ and pressed into glass discs. All oxides were measured in duplicate on a Siemens MRS-400 simultaneous spectrometer. Analyses that did not give summations between 99.4 and 100.3 percent were discarded and repeated. Estimates of the precision and accuracy of these data are given by *Rhodes* [1996] and *Rhodes and Vollinger* [2004]. Of specific relevance to this paper the precision for Al_2O_3 and MgO are 0.06 and 0.03 weight percent (2 sigma) respectively.

Because the differences in Al_2O_3 content of lavas from different Hawaiian volcanoes are small, I have restricted the study to data generated by X-ray Fluorescence Spectroscopy (XRF) in my laboratory at the University of Massachusetts, in order to minimize the possible effects of inter-laboratory bias. The data have been archived in the EarthChem Library and are available for download from www.earthchem.org/library/rhodes

3. The Al_2O_3 Content of Hawaiian Shield-Stage Lavas

Plots of major element compositions against MgO in Hawaiian shield-stage tholeiites show that at around 7% MgO there are two contrasting trends. In one trend, compositions range in MgO content between 6.8 and 35.0 wt. %. This is controlled primarily through olivine fractionation and accumulation, and plots along strong linear trends, the so-called “olivine-control trends”, that converge towards olivine compositions [e.g. *Powers*, 1955; *Murata and Richter*, 1966; *Wright*, 1971; *Wilkinson and Hensel*, 1988; *Clague et al*, 1995; *Rhodes*, 1995]. The other trend has cooled and crystallized beyond olivine control. These lavas are more evolved, with $\text{MgO} < 6.8\%$, and are multiply-saturated, crystallizing clinopyroxene and plagioclase in addition to olivine and sometimes orthopyroxene. Most analyses of Hawaiian tholeiitic glass fall into this category [e.g. *Garcia et al.*, 1989; *Moore and Clague*, 1992; *Clague et al*, 1995; *Stolper et al.*, 2004]. A few

highly evolved lavas exhibit compositional evidence for the crystallization of magnetite/ilmenite and even apatite.

In order to evaluate the Al_2O_3 content of lavas from Hawaiian tholeiitic shield-stage volcanoes I have restricted the study to lavas along olivine-control trends with $\text{MgO} > 6.8\%$. This avoids the effects on Al_2O_3 relative to MgO of plagioclase and clinopyroxene crystallization in evolved or mixed magmas. I have also confined the study to relatively unaltered lavas indicated by low Loss on Ignition ($\text{LOI} < 1\%$) and with $\text{K}_2\text{O}/\text{P}_2\text{O}_5 > 1.0$ [e.g. *Lipman et al.*, 1991; *Frey et al.*, 1994]. This study is based on over 1,800 major element analyses of Hawaiian shield-building tholeiites that lie on olivine-control trends. Most are from Mauna Loa, Mauna Kea and Kilauea.

3.1 Al_2O_3 in “Loa Trend” Volcanoes

3.1.1 Mauna Loa - the recorded magmatic history for Mauna Loa spans around 600 - 700 ka. Sub-aerial samples account for about 120 ka of this history and include data for historical lavas (1843-1984 [*Rhodes*, 1989; *Rhodes and Hart*, 1995]), prehistoric ^{14}C -dated lavas (0.2-36 ka [*Lockwood*, 1995; *Trusdell and Lockwood*, in press]) and samples from the Hawaii Scientific Drilling Project (0.3-120 ka [*Rhodes*, 1996; *Sharp et al.*, 1996; *Rhodes and Vollinger*, 2004; *Sharp and Renne*, 2005]). Submarine lavas from Mauna Loa’s southwest rift zone, and from the 1600 m high Kahuka landslide scarp that dissects the submarine southwest rift zone [*Garcia et al.*, 1995b], have $^{40}\text{Ar}/^{39}\text{Ar}$ ages from 59 to around 600 - 700 ka [*Jicha et al.*, 2012]. Recently, *Rhodes* [2015] utilized the relationship between water depth and $^{40}\text{Ar}/^{39}\text{Ar}$ ages to infer ages for other submarine Mauna Loa lavas. Despite this long magmatic history (~600 - 700 ka), Fig. 1a shows that Al_2O_3 vs MgO plots for these samples are extremely coherent and that there is no difference in Al_2O_3 at a given MgO content, either with age or between sub-aerial and submarine

lavas. Table 1 provides the regression data for these trends, confirming that there is no significant difference between them. Note that the enigmatic alkaline submarine Mauna Loa lavas erupted from submarine radial vents near Kealakekua Bay [Wanless *et al.*, 2006] are excluded from Fig. 1a because they plot off the olivine control trends.

3.1.2 Loihi - the data for Loihi come from samples discussed in *Garcia et al.* [1995; 1998]. All are tholeiitic and, with the exception of five samples that erupted in 1996, are of uncertain age. Although Loihi is isotopically a “Loa trend” volcano, and widely considered an end-member plume- source component [e.g. *Staudigal et al.*, 1984; *Garcia et al.* 1995; *Abouchami et al.*, 2005; *Weis et al.*, 2011], the major element compositions of its lavas are low in SiO₂ and high in CaO and TiO₂, similar to “Kea trend” lavas. Also, at a given MgO content, the Al₂O₃ content of Loihi lavas are slightly lower than other “Loa trend” lavas, closer to “Kea trend” lavas (Table 1; Fig. 1b).

3.1.3 Hualalai - analyses of the Hualalai shield-stage tholeiites were made on sub-marine lavas sampled by *Garcia et al.*, [1989]. Compositionally, they are indistinguishable from Mauna Loa lavas, with comparable Al₂O₃ at a given MgO (Fig. 1c, Table 1). On the other hand, post-shield Hualalai lavas are higher in Al₂O₃ at a given MgO than corresponding shield-stage lavas (Fig. 1c Table 1). This is consistent with post-shield Mauna Kea and Kohala lavas that also tend to be higher in Al₂O₃ at a given MgO than their shield-stage counterparts.

3.1.4 Mahukona - the data for Mahukona, the smallest, somewhat enigmatic, Hawaiian shield volcano are from *Garcia et al.* [2012]. They include both normal tholeiites and samples that are transitional to alkali basalt, probably post-shield. The tholeiites have Al₂O₃ values at a given MgO content that overlap with shield-stage tholeiites of Mauna Loa and Hualalai, whereas the

transitional basalts have higher Al_2O_3 values relative to MgO content, similar to post-shield lavas from Hualalai (Table 1, Fig. 1e).

3.1.5 Koolau - the data for Koolau come from *Frey et al.* [1994] and from the Koolau Scientific Drilling Project [*Haskin and Garcia*, 2004] and include samples from both the Makapuu- and the underlying Kahila-stages. The Makapuu-stage lavas of Koolau have generated a great deal of interest because of their high SiO_2 content relative to other Hawaiian tholeites and their end-member isotopic compositions with low Pb and Nd isotopic ratios and correspondingly higher Sr ratios [e.g *Frey et al.*, 1994; *Fekiacova et al.*, 2007; *Weis et al.*, 2011], characteristics widely inferred to indicate recycled oceanic crustal components, including sediments, in the plume source [e.g. *Hauri*, 1996; *Blichert-Toft et al.*, 1999; *Takahashi and Nakamura*, 2002; *Huang and Frey*, 2005; *Fekiacova et al.*, 2007]. In contrast, the Kahila-stage lavas are compositionally and isotopically similar to those of Mauna Loa [*Haskin and Garcia*, 2004; *Huang and Frey*, 2005; *Fekiacova et al.*, 2007]. Many of the Makapuu-stage lavas were rejected for this study because they were altered ($\text{K}_2\text{O}/\text{P}_2\text{O}_5 < 1.0$), or had fractionated beyond olivine-control ($\text{MgO} < 6.8\%$). Both Makapuu- and Kahila-stage lavas have Al_2O_3 values at a given MgO content that overlap with those of Mauna Loa and Hualalai, but tend to be slightly on the high side (Table 1, Fig. 1d).

3.1.6. Kahoolawe - the data are from *Huang et al.* [2005]. Samples from this volcano have high SiO_2 contents similar to Makapuu-stage Koolau lavas and broadly similar isotopic characteristics [*Huang et al.*, 2005]. Al_2O_3 values at a given MgO are also similar to, and overlapping, the Maakapuu-stage Koolau lavas (Table 1, Fig. 1d).

3.2 Al_2O_3 in “Kea Trend” Volcanoes

3.2.1 Kilauea - the data for Kilauea span at least 24 ka of its eruptive history and include historical lavas (1790 - 2008) [Pietruska and Garcia, 1999; Garcia et al, 2000; Marske et al., 2007], prehistoric ^{14}C -dated lavas that are mostly < 5 ka [Holcomb, 1987; Rhodes et al., 1989; Moore and Morris, 1996], lavas from the Hilina Pali over 22 ka in age [Chen et al., 1996] and submarine pillow lavas from the Puna Ridge [Clague et al., 1995]. On the basis of the thickness of palagonite rims on glass, Clague et al. [1995] speculate that the range in age of these submarine flows is around 0.7 - 24 ka. In spite of the age range of over 24 ka, and the fact that isotopic data indicate different proportions of source components during this period [e.g. Chen et al., 1996; Pietruszka and Garcia, 1999; Abouchanii et al., 2005; Marske et al., 2007; Weis et al., 2011] that fluctuate within decades [Pietruszka and Garcia, 1999], there are no discernible differences in Al_2O_3 at a given MgO content (Fig. 2a). Table 1 gives the regression data for these trends. There is a small difference in slope for the Puu Oo lavas, probably because the trends for these lavas are largely determined by olivine fractionation, whereas the others involve both olivine fractionation and accumulation. Nonetheless, the Al_2O_3 at a given MgO content overlaps with those of other Kilauea lavas (Table 1).

3.2.2 Mauna Kea - data used here for Mauna Kea comes primarily from the Hawaii Scientific Drilling Project (HSDP) [Stolper et al., 1996; Rhodes, 1996; Rhodes and Vollinger, 2004; Stolper et al., 2009; Rhodes et al., 2012]. Shield-stage lavas from Mauna Kea range in age from the transition to post-shield lavas at around 330 ka to extrapolated ages of around 690 ka at the bottom of the hole [e.g. Sharp et al., 1996; Sharp and Renne, 2005; Rhodes et al., 2012]. During this ~360 ka interval Mauna Kea erupted four distinct, and inter-layered, magma types [e.g. Rhodes and Vollinger, 2004; Stolper et al., 2004]. High-silica tholeiites (the type-1 lavas of Rhodes and

208 *Vollinger*, 2004) are the most abundant, erupted throughout the entire ~360 ka stratigraphic
 209 column, and including sub-aerial flows, submarine pillow lavas and hyaloclastites. In the upper
 210 part of the column (353 - 1740 mbsl) there are low-silica tholeiites (the type-2 lavas of *Rhodes and*
 211 *Vollinger*, 2004), inter-layered with the high-silica tholeiites. They occur as sub-aerial flows and
 212 sub-marine pillow lavas. Both magma types have similar incompatible element abundances and
 213 ratios, but both include lavas with different Pb isotopic ratios (the low-8 and mid-8 Pb isotopic
 214 groups of *Eisele et al.*, 2003). This indicates that the difference in SiO₂ between these two lava
 215 types is largely dependant on the depth of melt extraction and not on the extent of melting or the
 216 source composition [*Rhodes et al.*, 2012]. Between 1765 and 1810 mbsl there are enigmatic and
 217 controversial high - CaO hyaloclastites [*Stolper et al.*, 2004]. These are low- silica tholeiites, but
 218 have higher incompatible element abundances and are distinct from other shield-stage lavas.
 219 *Herzberg* [2006] argues, on the basis of their high CaO content, that they are the only Hawaiian
 220 magmas derived from a peridotite source, whereas *Rhodes et al.* [2012] infer that they are low-
 221 degree melts (similar to post-shield magmas) produced during a brief waning phase of Mauna
 222 Kea's shield-stage volcanism. In the lower part of the hole, below 2467 mbsl, pillow lavas and
 223 intrusive dikes of an additional low-silica tholeiite (the type-3 lavas of *Rhodes and Vollinger*,
 224 2004) are inter-layered with the high-silica tholeites. Geochemically and isotopically these are
 225 very similar to low-silica tholeiites from Loihi [*Garcia et al.*, 1995a; *Eisele et al.*, 2003; *Rhodes*
 226 *and Vollinger*, 2004; *Stolper et al.*, 2004].
 227 Despite the long ~360 ka magmatic record and geochemical and isotopic diversity indicative of
 228 different proportions of at least four source components, different depths of melt segregation and
 229 melt production, there are only small differences in Al₂O₃ at a given MgO content (Fig. 2b). There

is essentially no difference between the low- and high- silica (types-1 and 2) lavas or the high CaO lavas. The low-silica, Loihi-like, lavas type-3) are slightly higher in Al_2O_3 at a given MgO content. Table 1 provides the regression data for these trends, showing that there is no significant difference between them. Post-shield lavas are slightly higher in Al_2O_3 at a given MgO, similar to post-shield lavas from Hualalai and other volcanoes (Fig. 2d, Table 1).

3.2.3 Kohala - analyses of the Kohala shield-stage tholeiites were made on sub-marine lavas sampled by *Garcia et al.*, [1989] and sub-aerial samples studied by *Feigenson et al.* [1983]. Only a few sub-aerial samples are considered because most are altered ($\text{K}_2\text{O}/\text{P}_2\text{O}_5 > 1.0$) or evolved with $\text{MgO} < 6.8\%$. Nonetheless, the nine samples available plot well within the range for other “Kea Trend” volcanoes (Fig. 2c; Table 1) with similar Al_2O_3 content at a given MgO. The few post-shield lavas surviving the screening have higher Al_2O_3 values ($\sim 1\%$) with MgO, similar to Mauna Kea post-shield lavas, and post-shield lavas of the other volcanoes.

3.2.4 Loa vs Kea Trends - Table 1 provides the regression data for each of the volcanoes discussed above, and the estimated Al_2O_3 contents at 10 and 16 weight percent MgO. There is considerable overlap both within and between volcanoes. The maximum difference in Al_2O_3 , between the lower Loihi trend and that of the upper Makapuu-stage of Koolau is only about 0.8 weight percent. Figure 3 compares Al_2O_3 content versus MgO for the combined “Kea” and “Loa” trends. Again, there is considerable overlap between the two trends, but the regression through the Loa trend data is slightly higher in Al_2O_3 , by about 0.2 weight percent, than that through the “Kea” trend data. This difference, although small, is statistically significant. It may not, however, necessarily reflect a higher Al_2O_3 content in “Loa” trend lavas, but may be, in part, an artifact of olivine accumulation. The two trends are numerically dominated by data from Mauna Loa and

Mauna Kea (Table 1). At high MgO Mauna Loa picrites contain olivines that are slightly more forsteritic (modal and maximum values of Fo_{90} and $\text{Fo}_{91.3}$ respectively) than those of Mauna Kea (modal and maximum values of Fo_{88} and $\text{Fo}_{90.5}$ respectively) [Garcia et al., 1995; Baker et al 1996; Garcia, 1996; Putirka et al., 2011; Rhodes, 2015]. This small difference in Fo content of accumulative olivines could contribute to a small apparent difference in Al_2O_3 content of the two trends.

4. Discussion

Current thinking about the nature of the Hawaiian plume recognizes that it is probably thermally and compositionally zoned [e.g. Watson and McKenzie, 1991; DePaolo and Stolper, 1996; Ribe and Christiansen, 1999; Farnetani and Hofmann, 2010], with well-documented chemical and isotopic heterogeneity [e.g. Hofmann and White, 1982; Staudigal et al., 1984; Lassiter and Hauri, 1996; Blichert-Toft et al., 1999; Abouchami et al., 2000; Huang and Frey, 2005; Weis et al., 2011]. These heterogeneities are thought to related to varying contributions of recycled oceanic crust [e.g. Hofmann and White, 1982; Lassiter et al., 1996; Hofmann, 1997; Blichert-Toft et al., 1999]. They may, or may not, reflect source components derived from lithological differences between peridotite and recycled oceanic crustal eclogite/pyroxenite [e.g. Hauri, 1996; Takahashi and Nakajima, 2002; Sobolev et al., 2005; Herzberg, 2006; Jackson et al., 2012]. Whatever the cause, as a volcano traverses the mantle plume, we might expect to find evidence for: differences in melt production over time at individual volcanoes; differences in melt production between volcanoes; differences in the depth of melt segregation within and between volcanoes; and differences in the proportion of source components, within and between volcanoes.

Experimental melting of peridotite and basalt that is in equilibrium with a peridotite assemblage can produce a wide range in major element melt compositions from alkalic basalts to picrites, with compositions controlled both by the extent of melting and the pressure (depth) of melt segregation [e.g. Kushiro, 1996; Walter, 1998; Longhi, 2002; Salters et al., 2002]. Additionally, it has been proposed that Hawaiian magmas may be produced by melting of lithologically distinct mixtures of peridotite and re-cycled oceanic crustal components (eclogite/pyroxenite) [e.g. Hauri, 1996; Takahashi and Nakajima, 2002; Sobolev et al., 2005; Herzberg, 2006; Jackson et al., 2012]. In this case we should also expect differences in major element compositions, as well as in trace element abundances and isotopic signatures. These expectations are reflected in differences in most major elements between volcanoes, especially those from the “Loa” and “Kea” trends [e.g. Wright, 1971; Garcia et al., 1989; Rhodes et al., 1989; Moore and Clague, 1992; Frey and Rhodes, 1993; Frey et al., 1994; Hauri, 1996; Jackson et al., 2012] and sometimes within the long-time record of an individual volcano [e.g. Quane et al., 2000; Haskin and Garcia, 2004; Rhodes and Vollinger, 2004; Stolper et al., 2004]. However, although peridotite melting experiments indicate that Al_2O_3 should also vary with the extent and depth of melting, I have shown above (Table 1; Figs. 1 - 3) that the Al_2O_3 content of Hawaiian shield-building tholeiites is almost constant at a given MgO . What is the reason for this conundrum, and what does it tell us about the production and source of these magmas? The following sections explore these questions.

4.1 Harzburgite Melting

Experimental melting of peridotites and basalts indicates that the high- SiO_2 Hawaiian magmas from both the “Loa” and “Kea” trend volcanoes, which are volumetrically predominant during the shield-building stage, are produced at low pressures (< 2.2 GPa), corresponding to depths of less

than 70 km and that they are in equilibrium with a harzburgite residue. They are not in equilibrium with garnet lherzolite at any pressure [e.g. *Eggins, 1992; Wagner and Grove, 1997; Herzberg and O'Hara, 2002; Maaloe, 2004*]. Figure 4a uses the *Herzberg and O'Hara (2002)* parameterization of harzburgite melting to explore the effects of both pressure and extent of melting on MgO - Al_2O_3 relationships in Hawaiian tholeiites. In this figure the melt fraction at pressures ranging from 1 to 5 GPa is between 0.2 to 0.5. The Al_2O_3 and MgO contents of these melts are influenced by both depth of melt production and the extent of melting. Both result in a decrease in Al_2O_3 in the melt with increasing MgO indicating that Al_2O_3 is incompatible during melting. On the other hand, at a given MgO concentration, melts generated at higher pressures tend to have slightly higher Al_2O_3 than those produced at a lower pressure. From these relationships, we should expect to see a significant range of compositions, including Al_2O_3 , for magmas produced by differing degrees of melt extraction and at different depths within the plume, or from ascending polybaric melt columns in different parts of the plume. This is inconsistent with the data for Al_2O_3 presented earlier.

4.2 Lherzolite Melting

I have utilized data from the melting experiments of *Walter [1998]* and those of *Longhi [2002]*, reported in *Salters et al., [2002]*, to evaluate the effects of the depth and extent of melting on the Al_2O_3 and MgO contents of melts from lherzolite (Fig. 4b). The experiments of *Walter [1998]* were on the West Kettle River spinel peridotite between 3 and 7 GPa. The composition of this peridotite is thought to be close to that of primitive mantle or pyrolite [*Walter, 1998*]. In contrast, the experiments of *Salters et al., [2002]* were on synthetic picritic melts in equilibrium with a peridotite assemblage between 1 and 3.5 GPa. Despite the differences in approach the results are

very consistent (see also *Putirka et al.*, 2011, Fig. 6c). There is an overall trend of decreasing Al_2O_3 and increasing MgO with depth of melting between 2 and 7 GPa (Fig. 4b). This trend is similar to, and overlaps with, the trend shown previously for harzburgite melting (Fig. 4a). At pressures below 3 GPa, where garnet is absent, the results are similar to those inferred for harzburgite melting (Fig. 4a), and both depth and extent of melting result in a significant decrease in Al_2O_3 with increasing MgO.

Melts generated at pressures above 3 GPa, where garnet is stable, behave quite differently. An overall marked decrease in Al_2O_3 and increase in MgO with increasing pressure prevails but, in contrast, increase in melt production at a given pressure results in a slight increase in Al_2O_3 with increasing MgO (Fig. 4b). This indicates that garnet, the main source of Al_2O_3 in the melt, is a residual phase at these pressures.. The overall trend of experimental melts overlaps the compositional trend for Hawaiian shield-building tholeiites at around 18 to 22 weight percent MgO at pressures between 4 and 6 GPa. *Putirka et al.* [2011, Fig. 6] show a similar overlap for Al_2O_3 and other oxides at these pressures. That this is close to the inferred MgO content of parental Hawaiian tholeiitic magmas, based on olivine - melt relationships, is surely no coincidence [e.g. *Wilkinson and Hensel*, 1988; *Rhodes*, 1995; *Norman and Garcia*, 1999; *Rhodes and Vollinger*, 2004; *Putirka*, 2005; *Putirka et al.*, 2011]. Additionally, because these melting trends are only slightly positive (Fig. 4b), there is little increase in Al_2O_3 irrespective of the range in melting. There is also a tendency for the slopes of the melting trends to increase with increasing pressure, implying a greater garnet influence with depth. This provides a ready explanation for the almost constant Al_2O_3 at a fixed MgO for all Hawaiian shield-building tholeiites.

4.3 Melting Models

Here, I further explore the consequences of melting of garnet-free and garnet-bearing peridotites on the MgO and Al₂O₃ contents of basaltic melts, in order to broadly simulate the experimental melting trends of *Walter* [1998] and *Salters et al.* [2002]. This is shown in Figure 4c using the simple non-modal batch melting equation of *Shaw* [1970].

$$C_o = C_s * [1/(D_o + F(1 - P_o))]$$

C_o is the concentration in the melt, C_s is the concentration in the source, F is the melt fraction, D_o is the bulk distribution coefficient ($D_o = F_{ol} * Kd_{ol} + F_{opx} * Kd_{opx} + F_{cpx} * Kd_{cpx} + F_{ga} * Kd_{ga}$) based on the modal mineralogy of the source and the distribution coefficients (Kd) between minerals and melt. P_o is based on the fraction of minerals (P) entering the melt and the distribution coefficients (Kd) between minerals and melt ($P_o = P_{ol} * Kd_{ol} + P_{opx} * Kd_{opx} + P_{cpx} * Kd_{cpx} + P_{ga} * Kd_{ga}$).

The distribution coefficients for Al₂O₃ and MgO between mineral and melt are obtained from the experimental data of *Walter* [1998] and *Salters et al.* [2002]. Considering the different experimental approaches, they are encouragingly coherent. Most correlate strongly with both temperature and pressure (Appendix A)

The MgO and Al₂O₃ content of the starting material (the West Kettle River spinel peridotite (KR 4003)) in these experiments, are 37.3 and 4.26 weight percent, respectively. The mineralogical modal proportions at different pressures are from *Walter* [1998, Table 8, Fig. 3]. It is important to note that at 3 GPa the melt-free peridotite contains only about 2 percent garnet, but the garnet content increases with increasing pressure up to around 17 percent at 7 GPa. Accompanying this increase there is an increase in the Kd for Al₂O₃ in garnet (Appendix A). D_o is calculated from these Kd's and the modal proportions and P_o from the Kd's and the melt reactions of *Walter* [1998,

Table 9] at 3 and 7 GPa. Additionally, in order to simulate the experimental melt trends I have empirically estimated D_o and P_o from compositions of successive experimental melts at a given P. The results of these calculations are shown in Figure 4c, showing melting curves at various pressures. Melting is shown between 1 and 30 percent in increments of 10 percent. Broadly speaking, these melting curves follow the experimental melting trends shown in Figure 4b. At 3 GPa the calculated melting curve conforms with the experimental trend and Al_2O_3 is incompatible, decreasing with increased melting and MgO. This is because the 2 percent garnet in the starting peridotite is rapidly consumed by melting. Note also that melting from 10 to 20 percent results in a decrease in Al_2O_3 from 13.6 to 12.7 percent accompanied by an increase in MgO from 18.5 to 19.1 percent. Importantly, at pressures above 3 GPa, Al_2O_3 is compatible in all melting curves, increasing slightly with increased melt production and increasing MgO. Consequently, the calculated melting curves also do a better job of mimicking the experimental trends, although the range in MgO is not as great. In particular, there is a progressive decrease in Al_2O_3 with increasing MgO as pressure increases. Of particular relevance to the problem of the constancy of Al_2O_3 at a given MgO in Hawaiian shield-building lavas, is the fact that between 4 to 7 GPa, with increased melting at a fixed pressure, both Al_2O_3 and MgO increase only slightly. For example, at 4 GPa, melting from 10 to 20 percent results in an increase in Al_2O_3 of only 9.8 to 9.9 percent for an increase in MgO from 18.6 to 19.0 percent. Similarly, at 4.5 GPa, melting from 10 to 20 percent results in an increase in Al_2O_3 of only 8.3 to 8.5 percent for an increase in MgO from 20.7 to 21.2 percent. Comparable results are obtained for models of aggregated fractional melts. Figure 4c allows us to qualitatively evaluate the possible role of polybaric melting. Small

amounts of melt formed at around 4.5 to 4.0 GPa will be on or close to the Hawaiian trend, but with larger degrees of partial melting at shallower depths (~ 3 GPa) the melting trajectory will depart from the Hawaiian trend and Al_2O_3 in the accumulated melts will be too high. The situation will be exacerbated by continuing melting at even shallower depths (~ 2 GPa). It appears, therefore, that melting is not polybaric, and that the melt composition is established at around 4 GPa. The situation is analogous to that described by *Matzen et al.*, (2013) where the partitioning of Ni between olivine and melt during melting appears to be established at pressures greater than 3 GPa rather than over a range of pressures as the magma ascends.

The answer to the conundrum of almost constant Al_2O_3 content at a given MgO, in Hawaiian shield-building tholeiites, irrespective of the history of the volcano or whether it belongs along the “Kea” or “Loa” trends, is I believe that melting takes place at sufficient depth, most probably at, or above, 4 GPa (> 120 km) and firmly within the garnet stability field. At these depths both Al_2O_3 and MgO are compatible, resulting in only small changes in concentration with significant changes in melt production. The situation is analogous to that of the abundances of Yb and Y in Hawaiian tholeiitic lavas. Both elements have high garnet/melt K_d 's. For example, *Salters et al.* [2002] report garnet/melt K_d 's for Y and Yb ranging between 1.6 to 2.4 and 3.0 to 5.1 respectively at 3.2 to 3.4 GPa. The garnet/melt K_d for Al_2O_3 is pressure-dependant (Appendix A) ranging from around 1.9 at 3 GPa to 3.7 at 7 GPa, intermediate between K_d values for Y and Yb. Both Y and Yb are almost constant at a given MgO in Hawaiian tholeiites, and Hofmann and co-workers [e.g. *Hofmann et al.*, 1984; *Hofmann*, 1997] convincingly argue that high La/Yb, but in particular almost constant Yb in MgO - normalized Hawaiian tholeiitic lavas, are indicative of an origin from a garnet-bearing peridotite source.

If the presence of garnet as a residual phase during melting is critical to the consistency of Al_2O_3 in Hawaiian shield-building tholeiites, it follows that the more garnet initially present in the peridotite, the greater the probability of it being a residual phase. Increasing pressure increases the stability and abundance of garnet in peridotite, but so also will relatively minor “fertilization” of peridotite by reaction with melts derived from crustal components entrained within the plume [e.g. *Yaxley and Green*, 1998] . There is a broad consensus that mantle plumes contain some re-cycled oceanic crust plus minor sediments [e.g. *Hofmann and White*, 1982; *Hauri*, 1996; *Hofmann*, 1997; *Blichert-Toft et al.*, 1999; *Huang and Frey*, 2005; *Fekiacova et al.*, 2007; *Jackson et al.*, 2012]. Exactly how much, and whether or not this re-cycled crustal component, occurs as significant, distinct lithologies, such as eclogite or garnet pyroxenite, and whether it, along with peridotite, is involved in the melting process, is a matter of vigorous debate [e.g. *Hauri*, 1996; *Norman and Garcia*, 1999; *Takahashi and Nakajima*, 2002; *Huang and Frey*, 2005; *Sobolev et al.*, 2005; *Herzberg*, 2006; *Putirka et al.*, 2011; *Jackson et al.*, 2012; *Rhodes et al.*, 2012; *Matzen et al.*, 2013]. It is not the purpose of this paper to get involved in this debate. However, if discrete bodies of eclogite or garnet pyroxenite co-exist with peridotite in an ascending plume, they will melt before peridotite, at greater depths in the plume [e.g. *Petermann and Hirschmann*, 2003; *Sobolev et al.*, 2005] and these melts will react with peridotite [e.g. *Yaxley and Green*, 1998], enhancing the clinopyroxene and garnet contents of the peridotite. Such minor “fertilization” of peridotites will also contribute to the constancy of Al_2O_3 during melting at shallower depths in the plume. It also follows that basaltic melts of secondary garnet pyroxenites, produced by extreme fertilization of peridotite [e.g. *Sobolev et al.*, 2005; *Herzberg*, 2006], will also be characterized by almost constant Al_2O_3 at a given MgO. It is beyond the

scope of this paper to evaluate such models. Suffice it to say, there are several recent publications arguing that these convoluted models are not required to produce Hawaiian basaltic magmas: peridotite is sufficient [e.g. *Putirka et al.*, 2011; *Rhodes et al.*, 2012; *Matzen et al.*, 2013; *Rhodes* 2015].

4.4 What about the SiO₂ conundrum?

If the almost constant Al₂O₃ at a fixed MgO requires melting of garnet peridotite at pressures around 4 GPa, we now have combined trace and major element evidence supporting the formation of Hawaiian shield-building tholeiites from garnet peridotite deep within the mantle, probably at depths around 120 km. Experimental melting of peridotite and basalts at these pressures shows that the resulting parental basaltic magmas have low SiO₂ contents (see excellent summaries by *Longhi*, 2002; *Stolper et al.*, 2004; *Putirka et al.*, 2011). For example, the experiments of *Walter* [1998] between 3 and 5 GPa, produced melts from 44.8 to 46.9 percent SiO₂. Similarly, above 3 GPa the melts of *Salters et al.*, [2002] have around 45.6 to 46.0 percent SiO₂. On the other hand, with the exception of Loihi [e.g. *Garcia et al.*, 1995a] and compositionally similar lavas erupted on Mauna Kea [*Rhodes and Vollinger*, 2004; *Stolper et al.*, 2004], the vast majority of inferred parental Hawaiian tholeiitic magmas, including those from the lower SiO₂ “Kea” trend volcanoes, have SiO₂ contents exceeding 48%. This is a pervasive problem in Hawaiian petrology [e.g. *Eggins*, 1992; *Takahashi et al.*, 1993; *Wagner and Grove*, 1998; *Takahashi and Nakajima*, 2002, *Stolper et al.*, 2004]. Experimental melting of basalts and peridotites show that high-SiO₂ parental magmas, that is, those with high enough MgO to crystallize Fo₈₈₋₉₁, are produced at low pressures (< 2.2 GPa) corresponding to depths of less than 70 km and are in equilibrium with a harzburgite residue [e.g. *Herzberg and O’Hara*, 2002; *Maaloe*, 2004]. They are not in

equilibrium with garnet lherzolite at any pressure.

In order to explain this paradox, a number of authors have appealed to melt reaction between ascending magma and overlying depleted harzburgite [e.g. *Eggins*, 1992; *Wagner and Grove*, 1997; *Stolper et al.*, 2004; *Rhodes*, 2015]. The basic concept is that low-SiO₂ melts, produced by melting garnet lherzolite at depths of 90 - 130 km within the plume, interact on ascent with overlying, depleted, residual harzburgite (the residue of prior melt extraction within the plume). The reaction involves the dissolution of orthopyroxene from the harzburgite, compensated for by the crystallization of olivine from the melt. As noted by *Stolper et al.* [2004], if melt - harzburgite reaction is a prevailing process in the Hawaiian plume, then high-SiO₂ magmas will be dominant and low-SiO₂ magmas will be the exception. *Wagner and Grove* [1997] demonstrated that the reaction of a low-SiO₂ melt with harzburgite to produce a Kilauea parental magma had little influence on the garnet diagnostic La/Yb. The only effect was to slightly decrease the abundance of both elements. Similarly, *Rhodes* [2015] showed that reaction of an inferred low-SiO₂ Mauna Loa parental magma (similar to Loihi) with residual harzburgite had little influence on the isotopic ratios of ⁸⁷Sr/⁸⁶Sr, ¹⁴³Nd/¹⁴⁴Nd and ²⁰⁶Pb/²⁰⁴Pb.

But what about Al₂O₃? Will reaction of a low-SiO₂ parental melt with overlying residual harzburgite influence the constancy of Al₂O₃ resulting from melting of garnet peridotite at greater depths? Using the reactants from table 4 of *Wagner and Grove* [1997], 4 GPa experimental melt compositions (19% MgO and 9.5% Al₂O₃) and olivine (Fo₉₀) from *Walter* [1998], and reactant 2.4 - 2.8 GPa orthopyroxene (31% MgO, 6.2% Al₂O₃) from *Salters et al.* [2002] results in a shallow reactant melt (15.1% MgO, 10.8% Al₂O₃) that is firmly located on the Hawaiian MgO vs Al₂O₃ trend (Fig. 5). *Stolper et al.*, [2004] suggest that clinopyroxene might also be a reactant in

addition to orthopyroxene. Figure 5 shows that dissolution of clinopyroxene should result in a slight decrease in Al_2O_3 , compensating from the slight increase attributed to orthopyroxene. The MgO content of this inferred melt (15%) is by design close to that of the most primitive picritic glasses yet found in Hawaii [Clague *et al.*, 1991]. Also, because of the relative uniformity of FeO in Hawaiian tholeiites, a magma with around 15% MgO should crystallize olivines of around Fo_{88-89} (assuming a $\text{KD}_{\text{Fe-Mg}}$ of 0.32) at the high end of the olivine phenocryst range (Fo_{87-91}) in Hawaiian picrites [e.g. Wilkinson and Hensel., 1988; Helz, 1987; Baker *et al.*, 1996; Garcia *et al.*, 1995; Garcia, 1996; Putirka, 2005].

In contrast with Al_2O_3 , Ni is highly compatible in both the crystallizing olivine and the reactant orthopyroxene in the melt - harzburgite reaction, and one might anticipate difficulties with the model. This is the position taken by Huang and Frey [2005] who argue that the melt - peridotite reaction cannot account for the differences in SiO_2 in Hawaiian magmas because there is not a significant difference in Ni concentrations between high and low SiO_2 lavas from Mauna Kea or between the Maakapuu and Kalihi stage lavas of Koolau. However, there is considerable uncertainty concerning: which k_d 's for olivine and orthopyroxene are appropriate [e.g. Beattie *et al.*, 1991; Putirka *et al.*, 2011; Matzen *et al.*, 2013]; the Ni content of the source peridotite (1900 - 2360 ppm); and the Ni content of the parental melts [e.g. Sobolev *et al.*, 2005; Putirka *et al.*, 2011; Rhodes *et al.*, 2012; Herzberg *et al.*, 2013). It is beyond the scope of this paper to evaluate these here, but this will be discussed in a subsequent publication.

There is a serious problem with both the melt reaction model, advocated here, and with the widely preferred alternative of simply melting peridotite at pressures < 2.2 GPa. Both must occur at lithospheric depths. Beneath Hawaii the lithosphere is thought to be 88 km to 110 km thick [e.g.

Priestly and Tilmann, 1999; Li et al., 2004]. Isotopic data, especially Os, Pb and Hf, preclude involvement of the Cretaceous lithosphere overlying the Hawaiian plume in the melting process [e.g. Hauri et al., 1996; Lassiter and Hauri, 1996; Blichert-Toft et al., 1999; Abouchami et al., 2000; Fekiacova et al. 2007]. Therefore, unless the depth estimates for melt segregation are unrealistic, both models require involvement of plume-derived materials. There are two possible explanations: the lithosphere has been thinned and seismic resolution is insufficient to distinguish between lithosphere and underlying buoyant depleted plume harzburgite; or plume derived melts and associated residual harzburgite intrude deep into the lithosphere [Phipps Morgan et al., 1995; Wagner and Grove, 1997].

In summary, this is a plausible explanation for maintaining the almost constant Al_2O_3 content of parental Hawaiian shield-building tholeiites by melting within a deep garnet-bearing peridotite plume source, followed by melt reaction, during ascent with residual harzburgite to produce the almost ubiquitous high- SiO_2 magmas that characterize shield-stage growth of Hawaiian volcanoes. The less common low- SiO_2 lavas of Loihi and Mauna Kea may have avoided reaction with the residual harzburgite by flow through olivine-lined channels formed by prior melt-harzburgite interaction [e.g. Kelemen et al., 1995; Stolper et al., 2004]. In the case of Loihi, at the leading edge of the plume, there is probably very little residual harzburgite available for reaction.

5. Conclusions

1. Study of over 1,800 major element analyses from my XRF laboratory of Hawaiian shield-building tholeiites shows that the variation in the Al_2O_3 content, at a given MgO content is

very small, about one weight percent at 10 percent MgO. In the case of volcanoes for which we have long magmatic histories (e.g. Kilauea, about 24 ka; Mauna Kea, about 360 ka; Mauna Loa, about 600 ka) there is no significant difference in Al_2O_3 over time, nor are there differences between subaerial and submarine lavas, or magma types with different SiO_2 content or isotopic ratios. There is substantial overlap in Al_2O_3 in lavas between volcanoes belonging to the isotopically distinct “Loa” and “Kea” trends, but “Kea” trend volcanoes tend to erupt lavas with slightly lower Al_2O_3 relative to MgO than those from the “Loa” trend. Nonetheless, the average difference between the two trends is only about 0.2 weight percent (Fig. 3). Post-shield lavas from both the “Loa” and “Kea” trends are distinctly higher in Al_2O_3 relative to MgO than the earlier, shield-stage lavas.

2. The constancy of Al_2O_3 at a given MgO is contrary to what we anticipate from melting in a thermally-zoned and heterogeneous mantle plume. There are significant differences of major elements, trace elements and isotopic ratios between individual volcanoes, and sometimes over the growth of an individual volcano. These differences are usually attributed to different proportions of source components in the plume, and to differences in melt production and segregation at different depths as individual volcanoes are transported by the northwest-moving Pacific plate over the plume.

3. The vast majority of Hawaiian shield-building tholeiites, including those from both “Kea” and “Loa” trend volcanoes, are high in SiO_2 content. Experimental melting studies of peridotites and basalts indicate that high- SiO_2 parental magmas are produced at relatively shallow depths (< 70 km) in the mantle, with a harzburgite residue. Under such conditions Al_2O_3 will be incompatible and varying melting conditions should result in a range in

Al₂O₃ content in the magmas. As shown here, this is clearly not the case! The most plausible explanation is that Al₂O₃ is not incompatible during melting. The Kd's for garnet/melt correlate with both temperature and pressure, ranging from 1.9 at 3 GPa to 3.7 at 7 GPa. These are similar to the garnet/melt Kd's for Y and Yb. This requires melting to take place in the presence of garnet, from a garnet peridotite source, at depths exceeding 90 km, probably around 120 km.

Because both Al₂O₃ and MgO are compatible during melting at these depths, there will be little variation in both oxides with varying amounts of melt production, giving rise to the remarkable uniformity of Al₂O₃ at a given MgO in Hawaiian shield-building tholeiites. If the small difference in Al₂O₃ between "Kea" and "Loa" trend lavas is significant (Fig. 3) it could imply that "Loa" trend magmas result from slightly greater amounts of melting than "Kea" trend magmas, in agreement with the slightly lower concentrations of incompatible elements in "Loa" trend magmas. On the other hand, the markedly higher Al₂O₃ content of post-shield lavas relative to shield-stage lavas probably reflects garnet-free melting at the shallower (< 3 GPa) and cooler margins of the plume.

4. Relatively constant normalized Yb, and other HREE, abundances have been used as a strong argument for a garnet peridotite source for Hawaiian shield-stage magmas. We now have Al₂O₃, a major element oxide, providing the same information. These conclusions are contradictory to those from high-pressure experimental melting of peridotites and basalts. These experiments show that high-SiO₂ tholeiites, volumetrically the most abundant during the shield-building stage, were never in equilibrium with garnet, and were probably produced at depths of 70 - 90 km. In order to reconcile these two

contrasting arguments it appears possible that a low-SiO₂ parental magma, produced by melting garnet peridotite at depths of 100 - 120 km, reacts with a buoyant, depleted, harzburgite residue from prior melting within the plume, as the magma ascends through the upper regions of the plume. The result is a high-SiO₂ tholeiite that retains the garnet signature of its deeper origins.

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Table 1: Regression Data for MgO and Al₂O₃ Relationships in Hawaiian Shield-Building Lavas.

Figure Captions

Figure 1. Variations of Al₂O₃ with MgO in “Loa” trend volcanoes: (a) Mauna Loa; (b) Loihi; Hualalai; Koolau and Kahoolawe; Mahukona. The regression line is for the Mauna Loa data. The

data have been screened for samples that are considered to be altered ($K_2O/P_2O_5 < 1.0$) or have undergone fractionation beyond olivine control ($MgO < 6.8\%$).

Figure 2. Variations of Al_2O_3 with MgO in “Kea” trend volcanoes: (a) Kilauea; (b) Mauna Kea (shield-stage lavas only); (c) Kilauea, Mauna Kea and Kohala shield-stage lavas combined; (d) Mauna Kea post-shield lavas. In figures (a) and (b) the regression lines are for Kilauea and Mauna Kea respectively, and in (c) and (d) all the Kea shield-stage data are combined. The data have been screened for samples that are considered to be altered ($K_2O/P_2O_5 < 1.0$) or have undergone fractionation beyond olivine control ($MgO < 6.8\%$).

Figure 3. Comparison of variations of Al_2O_3 with MgO in “Loa” and “Kea” trend volcanoes, with regression lines for both trends. The data have been screened for samples that are considered to be altered ($K_2O/P_2O_5 < 1.0$) or have undergone fractionation beyond olivine control ($MgO < 6.8\%$).

Figure 4. Variation of Al_2O_3 with MgO in peridotite and basalt melting experiments and in numerical melt modeling. (a) Parameterization of harzburgite melting at pressures ranging from 1 to 6 GPa from *Herzberg and O’Hara* (2002). Both increase in pressure and in melt production result in significant increases in MgO and corresponding decreases in Al_2O_3 . (b) Melting experiments on peridotite [*Walter, 1998*] and picrites [*Salters et al., 2002*] between < 3 and 7 GPa. Only melts produced by less than 50% melting are shown. The vectors of increasing melt production are shown for different pressures. At 3 GPa or less, increased melting results in increase in MgO and decrease in Al_2O_3 , similar to harzburgite melting. At pressures above 3 GPa both MgO and Al_2O_3 increase with increased melt production, indicating that garnet is a residual phase. Additionally, the slope of the melting trend tends to increase with increasing pressure

reflecting the greater stability of garnet at increased pressure. (c) Comparison of non-modal batch melting models with the experimental melting data summarized in (b). The melting intervals are at 1, 10, 20 and 30 percent. At 3 GPa or less, progressive melting results in increase in MgO and decrease in Al_2O_3 , and there is a significant difference in MgO and Al_2O_3 between melt intervals. Above 3 GPa both MgO and Al_2O_3 increase with progressive melting and the differences between melt intervals are very small. The curving bold arrow schematically shows a path of polybaric melt production between 4.5 and <3.0 GPa. See text for details. The experimental and modeling data are compared with the combined trend for “Loa” and “Kea” lavas in all three figures (small grey crosses).

Figure 5. Reaction of a 4 GPa garnet peridotite melt with harzburgite using the reaction of *Wagner and Grove* [1997, Table 4]. The olivine is an average 4 GPa olivine from *Walter* [1998] and the orthopyroxene and clinopyroxene are average 2.4 - 2.8 GPa compositions from *Salters et al.* [2002]. See text for details.

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Figure 1a

Figure 1b

Figure 1c

Figure 1d

Figure 2a

Figure 2b

Figure 2c

Figure 2d

Figure 3

Figure 4a

Figure 4b

Figure 4c

Figure 4d

Figure 5