

# **Along-strike Variation in Lavas of the Aleutian Island Arc: Implications for the Genesis of High Mg# Andesite and the Continental Crust**

Peter B. Kelemen<sup>1</sup>, Gene M. Yogodzinski<sup>2</sup> and David W. Scholl<sup>3</sup>

1: Dept. of Geology and Geophysics, Woods Hole Oceanographic Institution,  
Woods Hole, MA 02543; peterk@whoi.edu

2: Dept. of Geological Sciences, University of South Carolina, Columbia, SC 29208

3: Dept. of Geophysics, Stanford University, Stanford, CA, 94035;  
dscholl@pangea.stanford.edu

## ABSTRACT

Western Aleutian, high Mg# andesite compositions bracket the composition of the continental crust. This is not due to recycling of terrigenous sediments. Is juvenile continental crust being produced in the western Aleutians? If so, how? These questions form the rationale for our paper. Based on a compilation of whole rock geochemistry for approximately 1100 lava samples and 200 plutonic rock samples from the Aleutian island arc, we characterize along-strike variation, including data for the western part of the arc which has recently become available.

The composition of primitive magmas (molar Mg# > 0.6) varies systematically along the strike of the arc. Concentrations of SiO<sub>2</sub>, Na<sub>2</sub>O and perhaps K<sub>2</sub>O increase from east to west, while MgO, FeO, CaO decrease from east to west. Thus, primitive magmas in the central and eastern Aleutians (east of 174°W) are mainly basalts, while those in the western Aleutians are mainly andesites. Along-strike variation in Aleutian magma compositions may be related to a westward decrease in sediment input, and/or to the westward decrease in down-dip subduction velocity. <sup>206</sup>Pb/<sup>204</sup>Pb, <sup>207</sup>Pb/<sup>204</sup>Pb, <sup>208</sup>Pb/<sup>204</sup>Pb and <sup>87</sup>Sr/<sup>86</sup>Sr all decrease from east to west, whereas <sup>143</sup>Nd/<sup>144</sup>Nd increases from east to west. These data, together with analyses of sediment from DSDP Site 183, indicate that the proportion of recycled sediment in Aleutian magmas decreases from east to west.

Most proposed trace element signatures of sediment recycling in arc magmas do not vary systematically along the strike of the Aleutians, and do not correlate with radiogenic isotope variations. Thus, for example, Th/Nb and fractionation-corrected K concentration in Aleutian lavas are not related to the flux of subducting sediment. An exception is Th/La, which varies systematically along-strike and correlates with sediment flux and isotope ratios. Interestingly, Th/La is also strongly correlated with Ba/La, rendering it doubtful that Ba/La is a proxy for an aqueous fluid component derived from subducted basalt. Ce/Pb > 4 is common in Aleutian lavas west of 174°W, in lavas with MORB-like Pb, Sr and Nd isotope ratios. Ce/Pb > 4 is also observed in lavas from volcanoes behind the main arc trend in the central Aleutians. Thus, Ce/Pb in Aleutian lavas with MORB-like isotope ratios is not always low, and may be affected by a component derived from partial melting of subducted basalt in eclogite facies.

Enriched, primitive andesites, with high Sr/Y, steep REE patterns, and low Yb and Y concentrations, are an important lava type in the Aleutians west of 174°W. As proposed by Kay [1978], these have trace element characteristics consistent with their derivation by partial melting of subducted eclogite, followed by reaction with the overlying mantle during ascent into the arc crust. High Sr/Y and Dy/Yb, indicative of abundant garnet in the source of melting, are clearly correlated with major element systematics in Aleutian lavas, and indicate that lavas with a large proportion of "eclogite melt" component have high SiO<sub>2</sub>, Na<sub>2</sub>O and K<sub>2</sub>O. Thus, in the Aleutians, Na content in primitive lavas cannot be used as an unambiguous indicator of the degree of partial melting of mantle peridotite. However, the high SiO<sub>2</sub> of primitive andesites probably reflects the effects of equilibration between hydrous, alkali-rich melt and shallow mantle peridotite, and is only coincidentally similar to high SiO<sub>2</sub> in the original eclogite melt.

Enriched, primitive Aleutian andesites did not form as the result of crystal fractionation from primitive basalt, mixing of primitive basalt and evolved dacite, or partial melting of metasomatized spinel peridotite. Unless rutile is a stable residual phase during partial melting of garnet peridotite, they cannot be explained as the result of small degrees of partial melting of garnet peridotite, and, similarly, they probably are not formed by shallow reactions between ascending mantle melts and shallow peridotite that form garnet pyroxenite. Instead, all indications are that they are formed by partial melting of subducted eclogite, followed by reaction with the overlying mantle.

Most enriched, primitive andesites in the Aleutians are Miocene, but there is evidence that they are still produced at present in the western Aleutians. In previous work, it has been proposed that a partial melt of subducted

sediment is an important component in central Aleutian lavas. Thus, partial melting of subducted sediment and/or basalt is occurring beneath most of the present-day Aleutian arc. This is inconsistent with all recent thermal models for arcs. However, there is much evidence that these models systematically underestimate temperatures in parts of the mantle wedge. An alternative is that mantle viscosities are lower than has been assumed in dynamic models up 'til now, and that thermal convection in the wedge is more rapid than the subduction rate. If so, temperatures in subducted metabasalt and metasediment would rise above the hydrous fluid-saturated solidus, permitting partial melting if fluid is available.

If subducted sediment and/or basalt melt beneath the entire arc, why are enriched primitive andesites observed only in the west? We propose that this is due to the relatively cold mantle wedge in the west. In the east, a hotter wedge, fluxed by abundant hydrous fluids, gives rise to abundant, mantle-derived basalts which mix with and obscure the subduction zone melt component.

Partial melts of subducted sediment and basalt are not saturated with olivine, and will solidify during reaction with mantle peridotite at temperatures less than  $\sim 1100^{\circ}\text{C}$ . Porous flow is too slow to allow such melts to survive reaction with the overlying mantle. However, U/Th isotope systematics suggest that at least some enriched, primitive andesites are transported from source to surface at velocities of  $\sim 1$  m/yr, ruling out processes in which subduction zone melts solidify in the mantle, and then are transported by solid state flow to another part of the wedge where they remelt. Instead, the eclogite melt component in Aleutian lavas is probably transported through the cold base of the mantle wedge in fractures or in partially molten diapirs composed of hybrid melt + pyroxenite.

The trace element contents of liquids formed by reaction between small degree melts of eclogite and mantle peridotite are strikingly similar to the estimated composition of the continental crust. Similarly, major element contents of liquids produced by experimental reaction of silicic melt with mantle peridotite closely approach those of high Mg# andesites and continental crust, and of primitive andesites. Thus, we believe that reaction of melts of subducted eclogite with the mantle wedge is a viable model for continental genesis. In a parallel line of reasoning, many investigators have suggested that unusually orthopyroxene-rich mantle xenoliths derived from Archean cratons underwent  $\text{SiO}_2$  enrichment as a result of reaction between highly depleted mantle residues and rising partial melts of subducted eclogite. Thus, both continental crust and cratonic mantle peridotites may be related by a process of reaction between subduction zone melts and sub-arc mantle peridotites.

Enriched primitive andesites, partial melts of eclogite, and products of small amounts of reaction between eclogite melts and mantle peridotite under conditions of decreasing magma mass, all have middle to heavy REE slopes that are much steeper than those in typical Aleutian, calc-alkaline andesite and continental crust. Thus, direct partial melts of eclogite – without magma/mantle interaction – do not form an important component in the continental crust. Extensive reaction, with gradually increasing melt mass and melt/rock ratios  $\sim 0.1$  to  $\sim 0.01$ , is required to increase heavy REE concentrations to the levels observed in most calc-alkaline andesites and in continental crust.

At small melt/rock ratios, such as are required to explain REE systematics, compatible element concentrations and ratios (e.g., Mg#, Ni) in eclogite melts would be completely modified to values in equilibrium with mantle peridotite. As a result, the liquids parental to high Mg# andesites and continental crust are inferred to be primitive andesites with  $\text{Mg\#} > 0.6$ . Thus, in addition to the reaction of eclogite melt with mantle peridotite, an additional process is required to explain the relatively low Mg# of continental crust (0.45 to 0.55) compared to primitive andesite. We believe the most likely explanation is that primitive andesites fractionate ultramafic cumulates in the uppermost mantle or at the base of the crust, followed by “delamination” and return of these cumulates to the convecting mantle.

## 1. INTRODUCTION

In this paper, we present results of a compilation of whole rock geochemistry for approximately 1100 lava samples and 200 plutonic rock samples from the Aleutian island arc. Aleutian lava compositions have been compiled and analyzed in several previous studies (e.g., [Kay and Kay, 1994a; Kelemen, 1995; Myers, 1988]). We have combined these previous compilations, and added more recent data. The purpose of the present compilation is to characterize along-strike variation, including data for the western part of the arc, much of which has recently become available, and to evaluate current ideas about arc magma genesis in light of the spatial-geochemical patterns in the Aleutians.

We focus on evaluating hypotheses for the origin of andesites with high Mg/(Mg+Fe), or Mg#, which are abundant at and west of Adak at ~ 174°W (Figure 1). These lavas are important because they are similar in major and trace element composition to the continental crust. Such lavas are rare or absent in intra-oceanic island arcs other than the Aleutians. Also, the western Aleutians show the smallest influence of a subducted sediment component of any part of the Aleutians, so that it is unlikely that enrichments in elements such as U, Th, K and light rare earths are due to recycling of subducted, continental sediments.

Western Aleutian lava compositions bracket the composition of the continental crust. This is not due to recycling of terrigenous sediments. Is juvenile continental crust being produced in the western Aleutians? If so, how? These questions form the rationale for our paper.

Some results of this review are, of necessity, preliminary, because data are sparse for lavas in the Aleutians west of ~ 177°W (Figure 2). In fact, parts of the text of this paper were recently used in a research grant proposal for funding to support dredging of submarine volcanic cones between 177°W and 187°W (173°E). However, there is no telling if and when our proposed

dredging cruise will be funded. Meanwhile, we feel that it is timely to present preliminary results based on currently available data. We hope that the hypotheses advanced in this paper will provide a useful base for future investigations, not only in the Aleutians but in other arcs as well. For example, the along-strike variation in isotope ratios is systematic, and clearly related to variation in sediment input. Thus, we have been able to test various trace element “proxies” for sediment recycling in arc lavas. These proxies are in current use in studies of global geochemical cycling and in detailed investigations of the Central American and Marianas arcs, so our results are timely and of broad relevance.

## 1.1 Terminology

In describing lava compositions, we have used some terms that require definition. “High Mg# andesite” is defined simply as lava with 54 to 65 wt% SiO<sub>2</sub> and Mg# > 0.45. This could, in principle, include some “boninites”. However, classical boninites in the western Pacific generally have flat to light rare earth element (REE) depleted trace element patterns, whereas high Mg# andesites in the Aleutians are all light REE enriched. Note that MgO content plays no role in our definition of high Mg# andesite. Some Aleutian andesites with Mg# > 0.6 have MgO contents as low as 4 wt%. As a result, we have not used the familiar terms “high Mg andesite” or “magnesian andesite”.

Some Aleutian high Mg# andesites – particularly some lavas on Adak Island first reported by *Kay* [1978] – also have been called “adakites” (e.g., [*Defant and Drummond*, 1990]). The term adakite is used in a variety of contexts by different investigators, but generally refers to andesites and dacites with extreme light REE enrichment (e.g., La/Yb > 9), very high Sr/Y ratios (e.g., Sr/Y > 50), and low Y and heavy REE concentrations (e.g., Y < 20 ppm, Yb < 2 ppm). In the Aleutians, all lavas with these characteristics are high Mg# andesites. However, note that the definition of “adakite” outlined above does not specify a range of Mg#. In arcs other than the

Aleutians, many highly evolved lavas with low Mg# have been termed adakites. Thus, globally, not all adakites are high Mg# andesites. Similarly, most high Mg# andesites, in the Aleutians and worldwide, have  $La/Yb < 9$  and  $Sr/Y < 50$ , so most high Mg# andesites are not adakites. Finally, we note that for some workers “adakite” has a genetic connotation as well as a compositional definition. Some investigators infer that all andesites and dacites with extreme light REE enrichment, very high Sr/Y ratios, and low Y and heavy REE concentrations must include an important component formed via partial melting of subducted basalt in eclogite facies, and use the term adakite to refer to both lava composition and lava genesis interchangeably. While we believe that many “enriched andesites” may indeed include a component derived from partial melting of eclogite, we feel it is important to separate rock names, based on composition, from genetic interpretations. For this reason, we have not used the term “adakite” in this paper. Aleutian andesites and dacites with  $Mg# > 0.45$ ,  $La/Yb > 9$ , and  $Sr/Y > 50$ , plus  $Y < 20$  ppm and/or  $Yb < 1$  ppm, form an important end-member composition on most chemical variation diagrams. We refer to these as “enriched, high Mg# andesites”.

More informally, we have used the terms “tholeiitic” and “calc-alkaline”. For our purposes, the definitions of these terms proposed by either [Miyashiro, 1974] or [Irvine and Baragar, 1971] are approximately equivalent and equally useful. Similarly, we have informally used the term “primitive” to refer to lavas with  $Mg# > 0.6$ . This usage reflects our belief that lavas with  $Mg# > 0.6$  have undergone relatively little crystal fractionation, and are derived from a parental liquid that was in equilibrium with mantle peridotite with an olivine  $Mg# > 0.88$ . Some investigators have suggested that some lavas and plutonic rocks with  $Mg# < 0.6$  – globally and in the Aleutians – are direct partial melts of subducted basalts (e.g., [Hauri, 1996; Marsh, 1976; Myers, 1988; Myers *et al.*, 1985; Schiano *et al.*, 1997]). Indeed, this could be a viable hypothesis in some cases. However, it is very difficult to evaluate such hypotheses since many (most?) lavas with  $Mg# < 0.6$  have been affected by crustal differentiation processes. In this paper, we

have adopted the convention that lavas with  $Mg\# < 0.6$  are considered to be the products of differentiation from primitive magmas until proven otherwise.

Finally, for simplicity in data presentation, in this paper we have reported longitude in terms of degrees west of Greenwich so that, for example,  $170^{\circ}E$  is referred to here as  $190^{\circ}W$ .

## 1.2 Regional divisions

In presenting our results, we have used regional divisions that are somewhat different from those used in previous papers (Figure 1). For our current purposes, Aleutian volcanoes from Atka eastward are all rather similar, with the exception of Rechesnoi volcano on Umnak Island. In order to avoid the complications of potential contamination from continental crust and overlying sediments, we have not compiled data for volcanic centers in the “eastern Aleutians”, east of  $164^{\circ}W$ . The arc from  $164$  to  $174^{\circ}W$ , which we call the “central Aleutians”, has been studied extensively. Most of it is a classic, oceanic arc dominated by tholeiitic basalts and their differentiation products (e.g., [Kay and Kay, 1994a; Myers, 1988] and references cited therein). We call the arc west of Atka the “western Aleutians”. We have subdivided the western Aleutians into three regions: (1) “the Adak area” (volcanoes on Adak, Great Sitkin, Kanaga, and Bobrof at  $174$  to  $177^{\circ}W$ ), (2) “the distal Aleutians” from  $177^{\circ}W$  to  $187^{\circ}W$ , and (3) “the Komandorsky block” from  $188^{\circ}W$  to the Kamchatka Strait at  $\sim 195^{\circ}W$ , including Komandorsky and Medny Islands and nearby, submarine volcanoes.

These subdivisions of the western Aleutians serve to remind readers of some important compositional and tectonic distinctions. The Komandorsky block lies within a transcurrent plate boundary, and may currently be moving mainly with the Pacific plate rather than with North American plate [Ave Lallemand and Oldow, 2000; Geist and Scholl, 1992; Geist and Scholl, 1994]. For this reason, it is not entirely clear how the Komandorsky block is related to the modern Aleutian arc. However, as it turns out, there is a compositional continuum from the

Adak area, through the distal Aleutians, to the Komandorsky block. Adak and neighboring islands are heterogeneous, with both tholeiitic and calc-alkaline lavas (e.g., [Kay *et al.*, 1982; Myers *et al.*, 1985]). Lavas in the Komandorsky block are predominantly calc-alkaline andesites, though basalts are observed [Yogodzinski *et al.*, 1995; Yogodzinski *et al.*, 1994]. Distal Aleutian lavas are relatively homogeneous, compared to the Adak area, and are transitional between the dominantly tholeiitic, basaltic lavas of the eastern and central arc and primitive, calc-alkaline andesites and dacites with extreme trace element signatures that predominate in the Komandorsky block (Figure 3). As a result of this continuum in compositional variation, we presume that the Komandorsky lavas are indeed related to the rest of the western Aleutians. In fact, recent work supports this idea, and proposes that the Komandorsky block originated in the forearc region of the distal Aleutians, and later was transported ~ 700 km WNW along the transcurrent plate boundary [Rostovtseva and Shapiro, 1998; Scholl *et al.*, 2001].

### **1.3 Andesite paradox: Why does continental crust resemble arc lavas?**

A long-standing problem in Earth Science is to resolve the following, apparent paradox. It is well documented that the continental crust has a bulk composition very similar to high Mg# andesite lavas and plutonic rocks in subduction-related magmatic arcs, with molar Mg/(Mg+Fe), or Mg#, from 0.45 to 0.54 at 57 to 65 wt% SiO<sub>2</sub>, 8000 to 24,000 ppm K, La/Yb of ~ 5 to 20, and so on (see Figures 4, 5 and 6, and [Christensen and Mooney, 1995; Kelemen, 1995; Rudnick, 1995; Rudnick and Fountain, 1995] for reviews). High Mg# andesites (again, lavas with 54 to 65 wt% SiO<sub>2</sub> and Mg# > 0.45) occur almost exclusively in arcs [Gill, 1981]. Thus, it is commonly inferred that continental crust formed mainly as a result of arc processes (e.g., [Kay and Kay, 1991; Kelemen, 1995; Rudnick, 1995; Taylor, 1977], and many, many others). However, most currently active, oceanic arcs are dominated by basalts and low Mg# andesites rather than high Mg# andesite lavas (e.g., the Marianas arc, Figure 4C).

Various hypotheses have been proposed to resolve this paradox. (1) At some times and places, the net magmatic flux through the Moho beneath arcs has been andesitic rather than basaltic (e.g., [Defant and Kepezhinskas, 2001; Drummond and Defant, 1990; Kelemen, 1995; Kelemen et al., 1998; Martin, 1986; Martin, 1999; Ringwood, 1974]). (2) High Mg# andesites in arcs form as a result of intra-crustal crystal fractionation and magma mixing, involving mantle-derived basalts and their differentiates. Later, a dense mafic or ultramafic plutonic layer delaminated and was returned to the mantle, leaving an andesitic crust (e.g., [Jull and Kelemen, 2001; Kay and Kay, 1991; Kay and Kay, 1993; Tatsumi, 2000]). (3) High Mg# andesites in arcs form via fractionation of ultramafic cumulates from primary, mantle-derived basalts. The ultramafic cumulates remain, undetected, below the seismic Moho in arcs and continents (e.g., [Fliedner and Klempner, 1999; Kay and Kay, 1985b]). (4) Arcs may not have been involved at all. Processes of intra-crustal differentiation and delamination have yielded a crustal composition whose resemblance to high Mg# andesites in arcs is coincidental (e.g., [Arndt and Goldstein, 1989; Stein and Hofmann, 1994]). In this paper, we will use data from the Aleutians to evaluate and extend hypotheses (1) through (3).

#### **1.4 Calc-alkaline lavas in the western Aleutians: Juvenile continental crust?**

In the Aleutian island arc, particularly in the western Aleutians, high Mg# andesite lavas with compositions similar to continental crust are abundant (Figure 4A, 5 and 6). Though similar lavas are also found in the Cascades, Baja California, Central America, the southern Andes, the Philippines, and SW Japan (e.g., [Defant et al., 1991; Defant et al., 1989; Hughes and Taylor, 1986; Luhr et al., 1989; Puig et al., 1984; Rogers et al., 1985; Tatsumi and Ishizaka, 1982]), these other localities are underlain by older, continental basement and/or sediment derived from continental crust. Light REE enriched, high Mg# andesite compositions are rare or absent in intra-oceanic, island arcs other than the Aleutians (e.g., the Marianas, Figure 4C). As a

consequence, the western Aleutians offer the best opportunity to study the genesis of high Mg# andesites – and, by analogy, the formation of juvenile continental crust – in a setting which is clearly free of contamination from older continental material.

Enrichment of arc lavas in some elements that are important in understanding the genesis and evolution of continental crust, such as K, U, Th, Pb and light rare earth elements (REE), is sometimes attributed to recycling of components from subducted, continentally derived sediments (e.g., [Elliott *et al.*, 1997; Hawkesworth *et al.*, 1997; Hochstaedter *et al.*, 1996; Miller *et al.*, 1994; Plank and Langmuir, 1993]). Isotopic and geological evidence suggests that sediment input to the source of arc lavas is common in the central Aleutians, but is absent or minimal in the west (Figure 7). Thus, if the processes that form *juvenile* continental crust in an oceanic arc can be documented anywhere on Earth, it is in the western Aleutians.

High Mg# andesites are rare in oceanic arcs worldwide. How could such lavas accumulate to form large volumes of continental crust? There are several possible answers. First, the genesis of high Mg# andesites in arcs may involve processes that were more common in the Precambrian than they are today. For example, perhaps high temperatures in subduction zones are required, and these were common in the Archean (e.g., [Martin, 1986]). Second, formation of abundant high Mg# andesite may be related to specific events. If high subduction temperatures are required, during the Phanerozoic these may have been most common in special circumstances such as “ridge subduction” (e.g., [Rogers *et al.*, 1985]), in which very young, hot oceanic crust is subducted. Third, hydrous andesite magmas become much more viscous than basalts when they degas at mid-crustal pressures, and may commonly form plutonic rocks rather than erupting as lavas [Kelemen, 1995], as is suggested by data from the Aleutians (Figures 4B and 5, plus [Kay *et al.*, 1990]). If so, the proportion of basaltic versus andesitic lavas is not indicative of the bulk composition of arc crust. These three factors, together or individually, may

explain how continental crust was formed as a result of processes which are rarely evident in present-day arc lavas.

### **1.5 Along-strike variation in convergence rate**

Systematic, along-strike variation in Aleutian lava compositions may be related to along-strike variation in the rate of “down dip” convergence between the Pacific and North American plates. Because of the arcuate shape of the Aleutians, convergence is nearly orthogonal to the trench in the eastern and central arc, and strongly oblique in the west. In addition, oblique convergence leads to strain partitioning in which absolute plate motions within the western arc are intermediate between those of the North American and Pacific plates [*Ave Lallemand and Oldow*, 2000; *Geist and Scholl*, 1992; *Geist and Scholl*, 1994]. As a result, the trench orthogonal convergence velocity, which is ~ 60 to 75 mm/year beneath the arc from Adak eastward, decreases to < 40 mm/year beneath the distal Aleutians and the Komandorsky block (Figure 8). (Orthogonal convergence rates *beneath* the arc have been projected from values of orthogonal convergence rate versus longitude along the trench [*Fournelle et al.*, 1994], using an empirical function relating the longitude of a particular volcanic center to the longitude at the trench updip along the plate convergence vector [*Engelbretson et al.*, 1985]).

In contrast to convergence rate, the age of subducting oceanic crust entering the Aleutian trench (> 40 Ma) is ~ 50 to 60 Ma and does not vary systematically along-strike [*Atwater*, 1989; *Geist et al.*, 1988; *Lonsdale*, 1988; *Scholl et al.*, 1987]. According to the plate reconstructions of Lonsdale, during subduction of the now dead Kula-Pacific spreading ridge (ceased spreading at ~ 43 Ma), the subducting crust was 28 to 40 Myr old, and at any given time it was progressively *older* to the west. For example, using Lonsdale's “simplest reconstruction” of the present down-slab position of the dead Kula-Pacific spreading ridge, using a down dip convergence rate of 65-

70 km/my, and considering that the subducted spreading ridge had a NE-SW (not an E-W) trend when it went into the trench, then:

At  $\sim 168^\circ\text{W}$  (Umnak Island), the dead spreading ridge was subducted about 15 Myr ago, when the youngest crust was  $\sim 28$  Myr old. At the present, the crust entering the subduction zone is anomaly 25, which has an age of 59 Ma.

At  $177^\circ\text{W}$  (Adak), the dead ridge was subducted about 12 Myr ago. At this time the youngest crust, at the extinct ridge, was  $\sim 31$  Myr old. Crust subducting to the east and west of this point was older. At the present, the crust entering the subduction zone is anomaly 24, which has an age of 56 Ma.

At  $180^\circ\text{W}$  (Amchitka), the spent ridge was subducted about 8 Myr ago. At this time the youngest crust was  $\sim 35$  my old. At the present, the crust entering the subduction zone is anomaly 24, which has an age of 56 Ma.

At  $187^\circ\text{W}$  (Attu), the old ridge was subducted about 3 Myr ago. At this time the youngest crust was  $\sim 40$  Myr old. At present, the crust entering the subduction zone is anomaly 22, with an age of 52 Ma.

Using Lonsdale's "more plausible" reconstruction, which considers age offsets across subducted fracture zones, then:

At Umnak, the spreading center at subduction had a thermal age of 32 Myr.

At Adak the spreading center at subduction had a thermal age of 35 Myr.

At Amchitka the spreading center at subduction had a thermal age of 35 Myr.

At Attu the spreading center at subduction had a thermal age of 39 Myr.

At any particular time, the lateral gradient in average slab temperature along the ridge was not large, and in general the hottest subducting crust was in the central Aleutians. Taking the middle Miocene ( $\sim 10$ - $12$  Ma), for example, when lots of plutons were emplaced and many enriched, high Mg# andesites were erupted, the thermal age of the segments of the spreading

center entering the trench at and east of Adak was roughly 32 Myr. At the same time, west of 180°W, the thermal age of the crust entering the subduction zone was on the order of 49 Myr.

Convergence rate could affect lava composition in a variety of ways. With decreasing convergence rate, the down dip flux of subducted sediment is smaller (Figure 9). Slow convergence leads to both increased conductive cooling of the convecting mantle wedge beneath an arc and increased heating of the subducting plate (e.g., [Kincaid and Sacks, 1997], their Figure 10). There are many possible consequences of these thermal effects: (1) dehydration reactions in the subducting plate might occur at shallower depths beneath the forearc, limiting the amount of H<sub>2</sub>O available to flux melting in the mantle wedge, (2) anatexis of subducted material might begin at shallower depths, extend to higher degrees of melting, and/or become more common or extensive, (3) there might be a lower extent of melting in the colder mantle, and (4) cooling of the shallow mantle and lower crust could lead to crystal fractionation at higher pressure, perhaps within the shallow mantle below the seismic Moho. Assuming that volcanic output is proportional to the magmatic flux through the Moho, it seems evident that some combination of (1), (3) and (4) limits the overall magma flux from the mantle to the crust in the western Aleutians. With the exception of Tanaga volcano, none of the emergent volcanoes in the western Aleutians are as large as the twenty largest volcanoes in the central Aleutians [Fournelle et al., 1994].

## **1.6 Along-strike variation in sediment thickness and composition**

Variation in Aleutian lava compositions might be related to the composition and volume of subducted sediment. Based on geophysical data, moderate size (20-40 km wide and as much as 6 km thick) prisms of tectonically accreted trench floor sediment form the lower part of the landward side of the Aleutian trench from 164 to 185°W. West of about 185°W, the wedge seems to rapidly dwindle in size and opposite Attu (188°W), the prism is quite small, perhaps

just a few km wide and thick. Geophysical imaging, DSDP drilling, and rock dredging, imply that the mass of the modern accretionary prism began to form in latest Miocene time, i.e., 5-6 Myr ago [McCarthy and Scholl, 1985; Ryan and Scholl, 1989; Scholl et al., 1987; Vallier et al., 1994; von Huene, 1986].

Mass balance calculations imply that, during the past 5 Myr or so, no more than 20 percent of the volume of trench-floor sediment swept into the Aleutian subduction zone has contributed to the mass of the accretionary prism [von Huene and Scholl, 1991]. Based on more recent geophysical and drilling data, compiled during the past decade, it is now supposed that no more than 15 percent of the mass of ocean floor sediment that enters the Aleutian subduction zone is retained in an accretionary prism [Scholl and Huene, 1998].

Thus, it is possible to approximate the flux of sediment beneath a given part of the arc as the product of the trench-orthogonal convergence rate (Section 1.5), and the thickness of sediment in the trench. Figure 9 illustrates the thickness and inferred composition of sediments in the Aleutian trench, based on [McCarthy and Scholl, 1985; Ryan and Scholl, 1989; Scholl et al., 1987; Vallier et al., 1994; von Huene, 1986] and our unpublished data.

About half of the sediment derived from erosion of the Aleutian arc itself is shed southward toward the Aleutian Trench, with the other half accumulating in the Bering Sea. Very little of this material appears to make its way to the trench floor because the bulk of the sediment pools in the Aleutian Terrace, which is underlain by a trough of arc-derived detritus, ash debris, and pelagic (mainly diatoms) sediment as thick as 3- 4 km. The basin of the Aleutian Terrace runs along virtually the length of the Aleutian Ridge.

In general, sediments in the trench are composed largely of turbidites composed of sediments shed off the Alaska Range and older sediments derived from the Chugach, Wrangell, and St Elias Mountains. The trench axis slopes westward continuously to about 180°W.

Sediments thicken gradually from east to west between 160 and  $\sim 172^{\circ}\text{W}$ . West of  $172^{\circ}\text{W}$ , they thin gradually from east to west, and then more abruptly from about 1 km thick at  $\sim 182^{\circ}\text{W}$  where the Rat Fracture Zone intersects the trench to just a few hundred meters at  $\sim 190^{\circ}\text{W}$ . Even further west they thicken again a result of turbidite sedimentation from Kamchatka. The thinning of the sedimentary section in the Aleutian trench from  $182^{\circ}\text{W}$  to  $190^{\circ}\text{W}$  is accompanied by an increasing proportion of pelagic to continentally-derived sediments.

The major variations in the thickness and relative proportions of different sediment types occur west of  $182^{\circ}\text{W}$ . Because the convergence vector is strongly oblique in the western Aleutians, any sediments beneath, e.g., Buldir volcano in the distal Aleutians ( $184^{\circ}\text{W}$ ) would have entered the trench at a longitude of  $\sim 179^{\circ}\text{W}$ . Sediments overlying the Pacific - North America plate boundary west of  $\sim 184^{\circ}\text{W}$  may not be subducted at all. Thus, there is little systematic, along-strike variation in the thickness or composition of sediment being subducted beneath the central and western Aleutian arc. If these estimates are correct, the main control on sediment flux beneath the Aleutian arc is the trench-orthogonal convergence rate.

## **2 ALONG-STRIKE VARIATION IN LAVA COMPOSITION**

In this part of the paper, we emphasize systematic east to west trends in lava composition. Previous studies of along-strike variation in the Aleutians have not emphasized systematic trends along the entire arc. Instead, spatial variation in lava composition has been attributed to (1) varying stress regimes within a series of tectonic blocks along the arc (e.g., [*Kay and Kay, 1994a; Kay et al., 1982; Singer and Myers, 1990*]), or (2) the availability of volatiles and fluid associated with subduction of thick sedimentary sections deposited over ancient oceanic fracture zones (e.g., [*Miller et al., 1992b; Singer et al., 1992a; Singer et al., 1992b; Singer et al., 1992c*]). In keeping with (2), it has long been noted that some of the largest Aleutian volcanic centers overlie subducted oceanic fracture zones [*Kay et al., 1982; Marsh, 1982a*].

An exception is the observation of Fournelle et al. [*Fournelle et al.*, 1994] that, in general, the size of Aleutian volcanoes decreases from east to west, which they ascribed to decreasing subduction rate and magma flux from east to west. Myers and co-workers noted that calc-alkaline lavas tend to be erupted from smaller volcanic centers, whereas tholeiitic lavas predominate at larger centers [*Myers*, 1988; *Myers et al.*, 1986a; *Myers and Marsh*, 1987; *Myers et al.*, 1985; *Myers et al.*, 1986b]. Thus, Fournelle, Myers and co-workers observed, or predicted, that the proportion of calc-alkaline to tholeiitic lavas in the Aleutians increases from east to west.

## **2.1 Along-strike variation in Mg# and MgO**

Figures 4 and 5 illustrate along-strike variation in major element composition of lavas in the Aleutians. Figure 3A shows the variation in Mg#. Although the data show considerable scatter, due mainly to crystal fractionation, the average Mg# for volcanic centers apparently increases from east to west. In the western Aleutians, where samples are few, most of them are primitive. This is fortunate, because we rely heavily on the composition of primitive lavas to infer the nature of mantle-derived melts in the Aleutians. (Note that primitive lavas can be affected by magma mixing and assimilation processes, which – if ignored – could lead to erroneous conclusions concerning primary magma compositions; such effects are discussed in Section 4.2 of this paper).

In Figure 4B, we show MgO contents in primitive lavas as a function of longitude along the arc. MgO contents decrease from east to west. Figure 10 shows variation of MgO content as a function of Mg# in the Aleutians. It is evident that primitive magmas in the Aleutians have a variety of MgO contents, ranging from 4 to 18 wt%, and that most of this variation is not due to crystal fractionation. As a result, it is inappropriate to constrain the nature of Aleutian primary

magmas by comparing concentrations of incompatible elements, such as Na, K or various trace elements, in lavas with a given MgO content.

One could, instead, correct incompatible element concentrations for the effects of fractionation by extrapolating to a given Mg#. However, because the slopes of variation trends for, e.g., Na<sub>2</sub>O and K<sub>2</sub>O versus Mg# (Figure 10) vary from one volcanic center to another in the Aleutians, comparisons of fractionation-corrected compositions of lavas with Mg# < 0.6 are fraught with peril. For this reason, we have chosen to concentrate on comparing uncorrected, incompatible element concentrations in primitive lavas with Mg# > 0.6.

The MgO content of mantle-derived magmas depends strongly on temperature. For example, many workers have developed geothermometers which depend on equilibrium Mg partitioning between olivine and liquid (e.g., [Gaetani and Grove, 1998; Roeder and Emslie, 1970] and references cited therein). For the most recent of these [Gaetani and Grove, 1998], we can use an olivine/liquid Fe/Mg Kd = 0.35 - 0.013 (mol% Na<sub>2</sub>O + K<sub>2</sub>O), based on the data of [Baker et al., 1996], and the assumption that 80% of all Fe is ferrous to estimate equilibrium olivine compositions for primitive Aleutian lavas, and then derive a temperature estimate.

Based on the assumptions outlined in the previous paragraph, estimated temperatures of equilibration between olivine and primitive lavas at 1 bar vary from ~ 1500°C for three picritic lavas with 16 to 18 wt% MgO, or ~ 1350°C for more common basalts with ~ 10 wt% MgO, to 1000°C for lavas with 4 wt% MgO (Figure 3C). Because Fe<sup>3</sup>/Fe<sup>2</sup> and water contents for Aleutian melts prior to possible crustal degassing, crystallization and eruption are not yet known, it is not possible to quantify magmatic temperatures more accurately. Probably, the low MgO, high Na<sub>2</sub>O + K<sub>2</sub>O, andesite lavas in the western Aleutians also include abundant H<sub>2</sub>O, whereas the primitive basalts in the eastern Aleutians have much less H<sub>2</sub>O. If so, then the total variation in Aleutian primitive magma temperatures is on the order of 300°C or more. Well-studied

volcanic centers in the Aleutians exhibit a  $\sim 200^\circ\text{C}$  range of estimated temperatures for primitive lavas which is not related to Mg#, and therefore appears to be a primary feature. This variability makes it difficult to be sure, but in general there seems to be a systematic, along-strike variation in magmatic temperature. For example, the maximum estimated temperature varies from  $\sim 1350^\circ\text{C}$  (or even  $1500^\circ\text{C}$ ) in the central Aleutians to  $\sim 1200^\circ\text{C}$  in the western Aleutians.

## 2.2 Along-strike variation in silica and alkali contents

Figure 3 illustrates along-strike variation in  $\text{SiO}_2$ ,  $\text{TiO}_2$ ,  $\text{Al}_2\text{O}_3$ ,  $\text{FeO}$ ,  $\text{CaO}$ ,  $\text{Na}_2\text{O}$ , and  $\text{K}_2\text{O}$ .  $\text{SiO}_2$  and  $\text{Na}_2\text{O}$  show a systematic increase from east to west. There may be a similar trend for  $\text{K}_2\text{O}$ , but it is less clear.  $\text{CaO}$ ,  $\text{FeO}$  and molar  $\text{Ca/Al}$  shows a systematic decrease from east to west.  $\text{TiO}_2$  and  $\text{Al}_2\text{O}_3$  show no clear trend. These trends are clearer if lava compositions are plotted versus trench-orthogonal convergence rate (Figure 8). This is because the central Aleutians show relatively little variation in both lava composition and convergence rate, and plot as a group when convergence rate rather than longitude is considered.

Plots of major element oxides versus Mg# for Aleutian lavas (Figure 10) can be used to evaluate the amount of variation which can be produced by crystal fractionation from a parent with an Mg# of, e.g., 0.7 to produce a derivative liquid with an Mg# of 0.6. It is clear from these plots that the east to west variation in Aleutian lava compositions is not the result of systematically increasing degrees of crystal fractionation from east to west.

The high alkali contents of primitive lavas in the western Aleutians probably arise as a result of either (a) small degrees of partial melting of the mantle, (b) variation in mantle source composition, possibly including a component derived from subducted metabasalt and/or metasediment, (c) combined crystal fractionation and melt/rock reaction in the shallow mantle, or (d) mixing of relatively alkali-poor, primitive basalt and alkali-rich, evolved dacites or rhyolites. These possibilities will be discussed further in Section 4. However, we note here that

the lack of systematic variation in  $\text{TiO}_2$  and  $\text{Al}_2\text{O}_3$  suggests that the variation of alkali element concentrations cannot be explained as the result of different degrees of melting of a common source composition, or different degrees of crystal fractionation from a common parental magma (i.e., a variety of primitive magma types are required).

High alkali and  $\text{H}_2\text{O}$  contents could explain the high  $\text{SiO}_2$  contents in primitive lavas in the western Aleutians. Increases in Na, K and H contents increase the size of the olivine primary phase volume, thereby increasing the  $\text{SiO}_2$  contents of magmas that can be in equilibrium with mantle olivine [Hirschmann *et al.*, 1998; Kushiro, 1975; Ryerson, 1985]. Like continental crust, Aleutian high Mg# andesites are quartz normative. After decades of debate, it remains controversial whether quartz-normative andesites can be formed by *partial melting* of mantle peridotites [Baker *et al.*, 1996; Falloon *et al.*, 1996; Hirose, 1997; Kushiro, 1990; Mysen *et al.*, 1974]. However, in our opinion, experimental data [Baker *et al.*, 1994; Tatsumi and Ishizaka, 1982] illustrate that quartz-normative, high Mg# andesites similar to those in the western Aleutians can *equilibrate* with mantle peridotite, provided they have sufficiently high  $\text{H}_2\text{O}$  contents. In Section 4, we discuss whether such liquids could have formed by small degrees of melting of mantle peridotite, or whether their high alkali contents arise via reaction of partial melts of subducted metabasalt and/or metasediment with mantle peridotite.

### **2.3 Along-strike variation in heavy, radiogenic isotope ratios**

As previously documented by, e.g., [Yogodzinski *et al.*, 1995], there are systematic along-strike variations in Pb, Nd and Sr isotope ratios in Aleutian lavas.  $^{206}\text{Pb}/^{204}\text{Pb}$ ,  $^{207}\text{Pb}/^{204}\text{Pb}$ ,  $^{208}\text{Pb}/^{204}\text{Pb}$  and  $^{87}\text{Sr}/^{86}\text{Sr}$  all decrease from east to west, whereas  $^{143}\text{Nd}/^{144}\text{Nd}$  increases from east to west (Figure 7).

In the central Aleutians, high Pb isotope ratios have been ascribed to recycling of Pb from subducted sediments into Aleutian lavas (e.g., [Class *et al.*, 2000; Kay *et al.*, 1978b; Kay and

*Kay, 1994a; Miller et al., 1994; Myers and Marsh, 1987; Singer et al., 1992a*] and references therein). Indeed, very high Pb isotope ratios ( $^{207}\text{Pb}/^{204}\text{Pb} \sim 19.2$ ) were reported for voluminous, terrigenous sediments recovered in DSDP Site 183 at 52.57°N, 161.20°W, SE of the eastern Aleutians [*Miller et al., 1994*]. However, pelagic and metalliferous sediments from just above the basaltic basement in Site 183 have Pb isotope ratios more similar to mid-ocean ridge basalts (MORB) [*Peucker-Ehrenbrink et al., 1994*]. In fact, some metalliferous sediments from elsewhere in the Pacific have Pb isotope characteristics indistinguishable from MORB. Metalliferous sediments are not abundant compared to continentally derived turbidites in DSDP Site 183, but they have high Pb concentrations, and the relative abundance of pelagic sediment to terrigenous turbidites on the subducting Pacific oceanic crust may increase westward. Thus, without further evaluation, Pb isotope ratios alone cannot be used as an unambiguous tracer for a recycled sediment component in Aleutian lavas.

Pb isotope ratios in Aleutian lavas are positively correlated with  $^{87}\text{Sr}/^{86}\text{Sr}$  and negatively correlated with  $^{143}\text{Nd}/^{144}\text{Nd}$  (Figure 11). In contrast, pelagic and metalliferous sediments have high  $^{87}\text{Sr}/^{86}\text{Sr}$  and low  $^{143}\text{Nd}/^{144}\text{Nd}$ . Like Pb, Sr is likely to be mobilized in both aqueous fluids and melts derived from subducted sediment. Thus, the correlation of Pb, Sr and Nd isotope ratios in Aleutian lavas suggests that only the Aleutian lavas with high Pb isotope ratios include a substantial, recycled sediment component. In support of this, Figure 9 shows that  $^{207}\text{Pb}/^{204}\text{Pb}$  in Aleutian lavas is correlated with trench-orthogonal sediment flux (the product of sediment thickness and trench-orthogonal convergence rate).

The unusually low Pb and Sr isotope ratios in lavas from the Komandorsky block deserve special mention. We are unsure about the reason for these low ratios. A few samples of MORB from the North Pacific have similarly low values, but they are rare. It would be of great interest

to obtain samples of lavas formed at the extinct Kula-Pacific spreading ridge to see if such depleted MORBs were common along that ridge, particularly near its western end.

#### **2.4 Along-strike variation in proposed indicators of sediment recycling**

As a result of the along-strike variation in trench-orthogonal sediment flux (Section 1.6 and Figure 9), and the related variation in radiogenic isotope ratios in lavas (Section 2.3 and Figure 7), the Aleutian arc represents an ideal testing ground for proposed relationships between the flux of trace elements in subducted sediment and their concentration in arc lavas. For example, Plank and Langmuir [*Plank and Langmuir, 1993*] demonstrated that there is a worldwide correlation between the flux of subducted K and Th, and the concentration of these elements in primitive arc lavas. Similarly, Elliott et al. hypothesized that a vector towards high La/Sm, Th/Nb, low Ba/La in lavas from the Marianas arc represented addition of a component derived from partial melting of subducted sediment [*Elliott et al., 1997*]. They contrasted this with a vector toward high Ba/La in Marianas lavas, which they inferred was due to addition of an aqueous fluid derived by dehydration of subducted basalt. Similarly, Hawkesworth et al. found a correlation between Th/La and isotopic indicators of a recycled sediment component in lavas from the Lesser Antilles [*Hawkesworth et al., 1997*]. Noting that Th is relatively immobile in aqueous fluid/solid equilibria, compared to silicate melt/solid equilibria, all of these groups suggested that the Th-enriched, sediment component must be a partial melt of subducted sediment, in the Marianas [*Elliott et al., 1997*], the Lesser Antilles [*Hawkesworth et al., 1997*], and worldwide [*Plank and Langmuir, 1993*].

Figure 12 shows the variation in K, K/Na, La, La/Sm, Ba, Ba/La, Ba/Th, Th, Th/Nb and Th/La in primitive lavas along the strike of the Aleutian arc. With the exception of Ba/La and Th/La, none of these variables show systematic variation along the arc. Th/La is correlated with Pb isotope ratios in Aleutian lavas (Figure 13A), so it probably is a good proxy for the presence

of recycled sediment, as proposed by Hawkesworth et al. Interestingly, Ba/La also shows an apparent correlation with Pb isotope ratios (Figure 13B).

## **2.5 Evaluation of proposed indicators of an aqueous fluid component from subducted material**

### ***2.5.1 Barium/Lanthanum***

As noted in the previous section, Elliott et al. [Elliott et al., 1997] and many other workers have proposed that Ba is mobile in aqueous fluids, whereas La, Nb and Ta are relatively immobile, so that Ba concentration in primitive lavas, and Ba/La, Ba/Nb and Ba/Ta in more evolved lavas, can be used to detect the presence of an aqueous fluid component derived from dehydration of subducted oceanic crust and/or sediments. Except for Ba/La, we see no systematic variation of these factors along the Aleutian arc (Figure 14).

Interestingly, as previously noted [Kay, 1980; Kay and Kay, 1994a; Yogodzinski et al., 1994], Ba/La is positively correlated with Th/La and Pb isotope ratios in Aleutian lavas (Figure 13). Aleutian lavas with high Th/La have high  $^{207}\text{Pb}/^{204}\text{Pb}$  (derived from sediment) and high Ba/La. Recycling of Pb and other components in the Aleutians cannot be clearly separated into a Ba-rich fluid component and a Th-rich sediment melt. In this way, the pattern of Ba/Th variation is strikingly different from many other intra-oceanic arcs worldwide (e.g., Elliott et al., *this volume*). The relatively high solubility of Ba and Pb in aqueous fluids, compared to relatively insoluble Th, is well-documented (e.g., [Ayers, 1998; Ayers et al., 1997; Brenan et al., 1996; Brenan et al., 1995a; Brenan et al., 1995b; Johnson and Plank, 1999]). In contrast, Ba, Pb and Th all behave incompatibly during melting. Thus, the correlation of Ba, Pb and Th concentrations in Aleutian lavas suggests that transport of all these elements from subducted crust into the mantle wedge is mainly via a silicate melt, not an aqueous fluid.

### ***2.5.2 Strontium/Neodymium***

Virtually all primitive lavas in the Aleutians have superchondritic Sr/Nd ratios, which shows no systematic variation along-strike. Elevated Sr concentrations relative to the REE, for example as indicated by Sr/Nd ratios greater than the chondritic ratio ( $\sim 17$ ; [Anders and Grevesse, 1989]), could be indicative of Sr addition to the arc magma source via an aqueous fluid, since experimental data indicate that Sr is more soluble in high P, high T aqueous fluids than the REE [Ayers, 1998; Ayers *et al.*, 1997; Brenan *et al.*, 1996; Brenan *et al.*, 1995a; Brenan *et al.*, 1995b; Johnson and Plank, 1999; Kogiso *et al.*, 1997; Stalder *et al.*, 1998; Tatsumi *et al.*, 1986]. Sr/Nd in primitive Aleutian lavas is not correlated with Ba/La, as might be expected if enrichments of Sr, Ba and Pb were all the result of aqueous fluid addition to the arc source. Alternatively, superchondritic Sr/Nd could indicate the presence of a component derived from cumulate gabbro in subducted, oceanic lower crust, since plagioclase-rich cumulates are enriched in Sr relative to the REE. And finally, trace element modeling (Section 4 and Figure 23) shows that partial melts of mid-ocean ridge basalt in eclogite facies may have super-chondritic Sr/Nd. Thus, we simply note that the explanation for elevated Sr/Nd in the Aleutians is uncertain.

### ***2.5.3 Cerium/Lead***

Miller *et al.* argued that unradiogenic Pb in the source of Aleutian arc magmas was transported from subducted basalts into the mantle wedge in an aqueous fluid [Miller *et al.*, 1994]. They suggested that Pb is relatively mobile in aqueous fluids, while Ce is relatively immobile, so that that Ce/Pb can be used to distinguish between (a) Pb transported in aqueous fluids with low Ce/Pb, and (b) Pb in partial melts of the Aleutian mantle and/or subducted MORB with Ce/Pb greater than 10. This hypothesis is supported by more recent experimental data on fluid/rock partitioning [Brenan *et al.*, 1995a]. Sediments have low Ce/Pb, so Miller *et al.* could not use Ce/Pb to distinguish between melt and fluid transport of Pb from sediments in

Umnak lava with radiogenic Pb isotopes. However, since Ce/Pb is low (less than ~ 4) in all lavas from Umnak Island, regardless of their Pb isotope ratios, Miller et al. concluded that the unradiogenic, MORB-like Pb isotope ratios in some Umnak lavas were due to transport of Pb in aqueous fluids derived from subducted basalt. This led to the much repeated aphorism, “sediments melt, basalts dehydrate”. On this basis, one could infer that Ce/Pb should be low in all Aleutian lavas, especially those with unradiogenic Pb isotope ratios.

Complicating this picture are the trace element models of possible igneous processes beneath arcs, in Section 4 and Figure 23. They show that Ce may be fractionated from Pb during partial melting of eclogite, and during reaction of melts with mantle peridotite, producing a range of possible Ce/Pb ratios in melts even where the source Ce/Pb is constant. The resulting Ce/Pb can be greater than or substantially less than in the original source of melting.

In the Aleutian data that we have compiled, Ce/Pb shows a weak, negative correlation with Pb isotope ratios, and shows systematic variation along-strike (Figure 15). All of the data discussed in this paragraph are for lavas whose Pb isotope ratios are known, and whose Pb concentration has been determined by isotope dilution. Ce/Pb is greater than 4 and ranges up to ~ 18 for all Aleutian lavas with  $^{207}\text{Pb}/^{204}\text{Pb}$  less than 15.5, all of which are west of Adak. On Adak, 6 of 15 lavas have Ce/Pb greater than 4, ranging up to ~ 15. In the central Aleutians, ~ 50 lavas have Ce/Pb less than 4, while three have Ce/Pb greater than 4, ranging up to ~ 7. All three of the central Aleutian lavas with Ce/Pb from 4 to 7 are from volcanoes behind the volcanic front; two from Amak Island and one from Bogoslof.

To summarize, unradiogenic Pb isotope ratios, reflecting Pb derived from subducted oceanic crust and/or the sub-arc mantle, are found in lavas with both low and high Ce/Pb ratios. In the western Aleutians, where Pb isotope ratios are lowest, a great majority of lavas have relatively high Ce/Pb, probably indicative of Pb transport in a melt rather than an aqueous fluid. In the

central Aleutians, high Ce/Pb together with low  $^{207}\text{Pb}/^{204}\text{Pb}$  are observed only in volcanoes situated behind the volcanic front. Other central Aleutian samples with low  $^{207}\text{Pb}/^{204}\text{Pb}$ , such as some of the Umnak lavas studied by Miller et al., have low Ce/Pb. These low Ce/Pb ratios *could* reflect transport of Pb from subducted basalt in an aqueous fluid, but are *also consistent* with Pb transport in a partial melt of subducted basalt (see Section 4 and Figure 23).

## **2.6 Along-strike variation in proposed indicators of an “eclogite melt” component**

Kay proposed that a few highly light REE enriched, heavy REE depleted, high Mg# andesites – from the base of Moffett volcano on Adak Island and dredged from the seafloor near Buldir Island – were produced by partial melting of subducted basalt in eclogite facies, followed by reaction of this melt with the mantle during transport from the subduction zone to the surface [Kay, 1978a]. Because we feel use of the term “adakite” for these lavas is problematic (see Section 1.1), we refer to these lavas as “enriched, high Mg# andesites”.

Kay noted that enriched, high Mg# andesites had such high Sr/Y, steep REE patterns, and low Yb and Y concentrations, that they seemed to require partial melting of a source with abundant garnet, in which Yb and Y were compatible and La and Sr were highly incompatible. The source must have contained little or no plagioclase, in which Sr is compatible and Yb and Y are highly incompatible. Many igneous processes can fractionate light REE from heavy REE whereas, among minerals that are abundant in metabasalt and mantle peridotite, only garnet can produce a strong fractionation of middle REE, such as Dy from heavy REE such as Yb. Thus, Dy/Yb is better than La/Yb as an indicator of abundant garnet in the source of arc magmas. Figure 16A shows that Sr/Y correlates with Dy/Yb in Aleutian lavas, reinforcing the idea that both high Sr/Y and high Dy/Yb reflect fractionation between melt and abundant, residual garnet. For further discussion of the origin of the high Sr/Y, Dy/Yb and La/Yb component in western

Aleutian lavas, and particularly whether it really requires a component produced by partial melting of subducted basalt in eclogite facies, please see Sections 4.5 and 4.7.

We can look for along-strike variation in the abundance of a component with abundant, residual garnet in our data compilation using La/Yb, Sr/Y, and Dy/Yb ratios in primitive lavas, all of which should be high in magmas including a substantial component produced by small amounts of partial melting of basalt in eclogite facies followed by reaction with mantle peridotite. Plots of these ratios versus longitude reveal little systematic variation (Figure 17), because moderate La/Yb, Sr/Y, and Dy/Yb ratios are found in lavas throughout the arc. However, the highest ratios are found in lavas at and west of Adak Island. This is more easily seen in plots of Sr/Y versus La/Yb and Dy/Yb, in which it is clear that highly enriched lavas are found only in the western Aleutians (Figure 16).

Figure 16 shows that the high Sr/Y, La/Yb, Dy/Yb component in primitive Aleutian lavas is enriched in SiO<sub>2</sub>, Na<sub>2</sub>O, K<sub>2</sub>O, Ce/Pb, and Th/Nb, and low in MgO and Ba/La, compared to the lower Sr/Y lavas in the eastern and central Aleutians. In many of these plots, lavas from Piip volcano in the Komandorsky block [Yogodzinski *et al.*, 1994] form an important, third endmember. The Piip lavas share the major element characteristics of the enriched, high Mg# andesites, and their low Ba/La ratios, but they lack high Sr/Y, La/Yb, and Dy/Yb. As pointed out by Yogodzinski *et al.* [Yogodzinski *et al.*, 1994], many of the characteristics of Piip lavas can be explained as the result of a small degree of partial melting of spinel peridotite.

It is noteworthy that Ba/La and Th/La are low in lavas with high Sr/Y, Dy/Yb and Ce/Pb, whereas Sr/Y, Dy/Yb and Ce/Pb are low in lavas with high Ba/La and Th/La (Figure 18). If high Ba/La and Th/La ratios are proxies for a component derived from partial melting of subducted sediment, and high Sr/Y, Dy/Yb and Ce/Pb are caused by addition of a melt of subducted

eclogite, these two components are clearly different in the Aleutians. Both end-members are present in the western Aleutians (mainly in the Adak area).

## **2.6 Proposed indicators of other melt components derived from subducted basalt**

Marsh, Myers, Brophy, Fournelle and their co-workers (e.g., [Brophy and Marsh, 1986; Myers *et al.*, 1986a; Myers *et al.*, 1985; Myers *et al.*, 1986b]) have proposed that more typical, low Mg#, high alumina basalt magmas in the Aleutians were also produced by partial melting of subducted basalt. Because these lavas do not have high Dy/Yb and Sr/Y, this group called upon diapiric ascent of partially molten, subducted basalt into the sub-arc mantle. Decompression within these diapirs would lead to garnet breakdown, and subsequent separation of the melt from its residue would produce high alumina basalt with the isotope signature of subducted basalt but without the trace element signature indicative of an eclogite melt. However, the hypothesis that high alumina basalts are partial melts of subducted basalts is controversial and difficult to verify, because high alumina basalts may also be produced by mantle melting [Bartels *et al.*, 1991] and by crustal crystal fractionation processes (e.g., [Baker and Egger, 1983; Kay *et al.*, 1982; Kelemen *et al.*, 1990b; Sisson and Grove, 1993]).

Similarly, Drummond and Defant (e.g., [Defant and Drummond, 1990; Drummond and Defant, 1990]), Rapp and Watson [Rapp and Watson, 1995; Rapp *et al.*, 1991] and others have inferred that some low Mg# andesites and dacites might be direct partial melts of subducted eclogite, which did not interact with mantle peridotite. Again, this hypothesis is difficult to verify for arcs worldwide because low Mg# andesites - even those with high Sr/Y and so on - could be produced by differentiation of primitive basalts or andesites, or could be produced by melting within thick, garnet-bearing crust. In the Aleutians such ambiguities do not arise, because the crust is relatively thin, and Sr/Y and Dy/Yb ratios in Aleutian lavas are restricted to high Mg# andesites, and are not found in any lavas with Mg# < 0.5.

In this paper, we have adopted the assumption that low Mg# magmas are produced by differentiation of primitive magmas which were in Fe/Mg equilibrium with residual mantle olivine, unless proven otherwise.

### 3 PLUTONIC ROCKS IN THE ALEUTIANS

Plutonic rocks in the Aleutians have been studied mainly on Unalaska Island in the east and in the Adak area (e.g., [Kay *et al.*, 1983; Kay *et al.*, 1990; Perfit *et al.*, 1980b] and references therein). There are also some small, hypabyssal intrusions in the western Aleutians and on the Komandorsky Islands (e.g., [Tsvetkov, 1991; Yagodzhinski *et al.*, 1995; Yagodzhinski *et al.*, 1993]). Though some of the most mafic Aleutian gabbros may be “cumulate” (products of partial crystallization, from which remaining liquid was later removed), the high K contents of intermediate to felsic plutonic rocks in the Aleutians suggest that most are close to magma compositions.

Light REE enriched, K-rich, high Mg# andesite compositions similar to continental crust are more common among Aleutian plutonic rocks than they are among lavas (Figures 4 and 5). Although Kay *et al.* [Kay *et al.*, 1990] noted that plutonic rocks in the Adak area were somewhat more K-rich at a given Mg# than those on Unalaska Island, these variations are small compared to the along-strike variation in lava compositions.

Plutonic rocks analyzed to date - all Tertiary - may reflect an earlier phase of Aleutian magmatism. However, there is no identified change in convergence rate, age of subducting plate, or sediment source and abundance, which would explain a change in magma composition [Atwater, 1989; Engebretson *et al.*, 1985; Lonsdale, 1988]. Instead, the plutonic rocks may represent hydrous andesite magmas that were emplaced in the mid-crust after degassing of H<sub>2</sub>O left them too viscous to erupt, as previously suggested by Kay *et al.* [1990] and Kay and Kay [1994]. In contrast, low viscosity, H<sub>2</sub>O-poor, basalts may erupt readily. Thus, the proportion of

high Mg# andesites among lavas may be less than their proportion within the entire crust. If true, this is very important, because it is commonly assumed that lava data can be used to estimate the proportions of the primary melts that form arc crust.

## **4 GENESIS OF WESTERN ALEUTIAN HIGH MG# ANDESITES**

### **4.1 Crystal fractionation from primitive basalt? No.**

*Kelemen* [*Kelemen*, 1995] reviewed experimental data on partial melting of natural basalt and peridotite compositions, with and without added H<sub>2</sub>O, at a variety of oxygen fugacities, over a range of pressures from 1 bar to 3.5 GPa. Very few of these experiments have produced high Mg# andesite liquid compositions, and even fewer have produced high Mg# andesites with K<sub>2</sub>O concentrations comparable to Aleutian high Mg# andesites. Since 1995, Kawamoto melted basalts at high, unbuffered oxygen fugacity and produced a series of melt compositions that were very similar to some Aleutian high Mg# andesites [*Kawamoto*, 1996]. However, the oxygen fugacities in those experiments were probably much higher than in most arc lavas (fO<sub>2</sub> within 2 log units of the Ni - NiO oxygen buffer [*Gill*, 1981]) according to Kawamoto (pers. comm., 1997).

In considering the experimental data, it is also worthwhile to recall that partial melting/crystallization experiments are performed in a “closed system”, in which melts continuously undergo Fe/Mg exchange with residual crystals. In fractional crystallization, which is often closely approximated in nature (e.g., [*Carmichael*, 1964]), residual crystals are removed so that melts evolve to lower Mg# at a given SiO<sub>2</sub> and K<sub>2</sub>O content compared to closed-system crystallization. Thus, the experimental data approximate an upper bound on the SiO<sub>2</sub> and K<sub>2</sub>O contents which can be achieved at a given Mg# via basalt crystallization. From this perspective, it is striking that only a few experimental melts fall into the range of Aleutian high Mg# andesite compositions, and the range of estimated continental crust compositions (two from Sisson and

Grove, one in which the  $fO_2$  buffer failed, and another at 965°C saturated in hornblende and magnetite, plus the high  $fO_2$  experiments of Kawamoto).

Since the composition of continental crust is clearly an average, and it is apparent that there are basaltic rocks and differentiated low Mg# andesites incorporated into the continental crust, these must be balanced by a significant proportion of andesitic rocks with Mg# higher than the average value, similar to primitive Aleutian andesites. In the Aleutians at and west of Adak, andesites with Mg# between 0.45 and 0.6 are spatially associated with primitive andesites with  $Mg\# > 0.6$ . Also, both andesites with Mg# between 0.45 and 0.6 and primitive andesites have similar trace element concentrations, which are distinct from trace element patterns in primitive basalts. Therefore, it is apparent that andesites with Mg# between 0.45 and 0.6 and primitive andesites are genetically related. No experimental liquids formed by partial melting of basalts approach the compositions of Aleutian primitive andesites. For these reasons, we believe that it is unlikely that western Aleutian high Mg# andesites are produced primarily via crystal fractionation from primitive basalts.

#### **4.2 Mixing of primitive basalt and evolved dacite? Not for enriched, primitive andesite.**

In the Aleutians, it has long been argued that both tholeiitic and calc-alkaline lava series have a common parental magma, a mantle-derived picrite with a nearly flat REE pattern (e.g., [Kay and Kay, 1985a; Kay and Kay, 1994a; Nye and Reid, 1986a]). In this view, the tholeiitic series evolves mainly via crystal fractionation, perhaps at relatively low  $H_2O$  fugacity, whereas the calc-alkaline series has a more complex origin. Some light REE enriched dacite magmas are produced by crystal fractionation involving hornblende, which fractionates middle REE from light REE. Then, mixing of these evolved dacites with primitive basalt produces light REE enriched, high Mg# andesites such as those on Buldir Island in the western Aleutians.

It is well documented that Aleutian magmas, and calc-alkaline andesites in general, commonly contain oscillatory zoned clinopyroxene crystals which could be indicative of magma mixing (e.g., [Brophy, 1987a; Conrad and Kay, 1984; Conrad et al., 1983; Kay and Kay, 1985a]). One possible explanation for both oscillatory zoning in clinopyroxene and high Mg# andesite lavas is mixing of primitive basalt with evolved dacite or rhyolite. Because the primitive basalt end-member has much higher Mg and Fe concentrations than the evolved end-member, mixing produces a concave downward hyperbola on plots of, e.g., Mg# versus SiO<sub>2</sub> or K<sub>2</sub>O. If the evolved end-member is also light REE enriched and heavy REE depleted, perhaps as a result of crystal fractionation in the crust, then a similar trend is produced on plots of Mg# versus La/Yb (Figure 19).

Such mixing processes, or very similar processes in which primitive basalt assimilates evolved, granitic rocks, have long been proposed to explain the origin of high Mg# andesites in general, and more specifically in the Aleutians (e.g., [Brophy, 1987a; Conrad and Kay, 1984; Conrad et al., 1983; Kay and Kay, 1985a] and references therein). Mixing of low viscosity basalt and high viscosity, siliceous liquids may be physically improbable [Campbell and Turner, 1985]. On the other hand, there are very well documented cases of assimilation processes that produce calc-alkaline andesites [Grove et al., 1988; McBirney et al., 1987]. Furthermore, in considering the genesis of continental crust, tectonic juxtaposition of basaltic and evolved rocks could produce an average with the composition of high Mg# andesite without any chemical mixing.

However, it is apparent that mixing of primitive basalt and evolved dacite or rhyolite cannot explain the origin of enriched, high Mg# andesites in the Aleutians at and west of Adak. As noted in Section 2.6, among primitive Aleutian magmas, high Sr/Y, La/Yb and Dy/Yb ratios are found only in andesites and dacites, never in basalts (Figure 16). Furthermore, crystallization of

plagioclase, in which Sr is compatible, lowers Sr/Y, so that all evolved Aleutian magmas, with  $Mg\# < 0.45$ , have  $Sr/Y < 60$ . Thus, in the Aleutians, neither primitive basalts nor evolved magmas have the trace element characteristics which distinguish the enriched, primitive andesites.

Yogodzinski and Kelemen [Yogodzinski and Kelemen, 1998] investigated the trace element contents of zoned clinopyroxene crystals in enriched, primitive Aleutian andesites, and compared them to trace element concentrations in clinopyroxenes from other Aleutian magma types. Primitive andesites contained clinopyroxenes with high Sr/Y, Dy/Yb and La/Yb, while primitive basalts contained clinopyroxenes with much lower ratios. Evolved lavas had relatively low Sr/Y, Dy/Yb and La/Yb, similar to the primitive basalts. In zoned clinopyroxenes in the primitive andesites, we found that the highest Sr/Y, Dy/Yb and La/Yb ratios were in the parts of crystals with the highest Mg#. Lower Mg# portions of the same crystals had lower Sr/Y, Dy/Yb and La/Yb, indicative of mixing of a primitive, highly enriched component with an evolved, less enriched component.

As Yogodzinski and Kelemen noted, trends of Aleutian lava compositions parallel those observed in the zoned clinopyroxene crystals. Primitive basalts and most evolved lavas - basaltic, andesitic, and even more silica rich - have relatively low Sr/Y, Dy/Yb and La/Yb, whereas enriched, primitive andesites have much higher Sr/Y, Dy/Yb and La/Yb (Figure 20). In fact, this bimodal distribution of trace elements is reflected in a similar, bimodal distribution of major element contents of primitive lavas (Figure 21). Among western Aleutian lavas, a mixing trend can be seen from enriched, primitive andesites toward less enriched trace element ratios in andesites with moderate Mg#. Thus, it is apparent that there are two primitive lava types in the western Aleutians. As previously emphasized (e.g., [Kay and Kay, 1985b; Kay and Kay, 1994a; Yogodzinski and Kelemen, 1998]), magma mixing has played an important role in the genesis of

many high Mg# andesites at and west of Adak, but the highest Mg# end-member in this mixing process was an enriched, primitive andesite, not a primitive basalt.

It is not clear whether calc-alkaline lavas and plutonic rocks with high Mg# andesite compositions in the central Aleutians include a component derived from enriched, primitive andesites, or whether, instead, they formed via mixing of primitive basalt with enriched, evolved dacite or rhyolite. In the Adak area, hornblende-bearing, dacitic tephra include glasses with La/Yb generally between 6.5 and 7.5, and one from Kanaga has La/Yb=8.9 [Romick *et al.*, 1992]. However, there are very few dacites and rhyolites in the central and eastern Aleutians with sufficiently high La/Yb (greater than ~ 10) to explain the light REE enrichment in high Mg# Aleutian plutonic rocks (Figure 4). In our database, only 5 of ~ 900 samples from the central and eastern Aleutians have more than 60 wt% SiO<sub>2</sub> and La/Yb > 7. Thus, it seems plausible that magma mixing is not the general cause of light REE enrichment in Aleutian plutonic rocks. Instead, an enriched, primitive andesite component may have been an important mixing end-member throughout the arc.

Apparently, even in the Komandorsky block and on Adak, where primitive lavas include both enriched andesites and more “normal” basalts, there has been very little mixing between these different, primitive magma compositions. Perhaps this is due to the very different viscosities of andesitic and basaltic liquids. The mixing trend observed between enriched, primitive andesites and more evolved, but less enriched andesites may be facilitated by their similar viscosities.

#### **4.3 An important role for recycled, continental sediment? No.**

Leaving aside the major element composition of Aleutian lavas for a moment, it might be proposed that the enrichments in Th, U, Pb, Rb, K, Sr, light REE, and so on, in western Aleutian lavas - different from other intra-oceanic arcs and similar to continental crust - is simply the

result of recycling of a continentally-derived sediment component. We believe that the isotopic and geologic evidence clearly indicate that this is not the case (see Sections 1.6 and 2.3).

Although the concentration of some elements in primitive arc magmas worldwide may be related to the flux of these elements in subducted sediments [*Plank and Langmuir, 1993*], this is not the case in the Aleutians, at least not at and west of Adak Island.

#### **4.4 Partial melting of spinel peridotite metasomatized by aqueous fluid? No.**

As noted in Section 1.5, the relatively slow, trench orthogonal convergence rate in the Aleutians west of Adak probably results in a relatively cold mantle wedge, due to conductive heat loss from the mantle to the subducting plate. Also, slow convergence combined with relatively young subducting crust will lead to a relatively hot subduction zone. In such a setting, most metamorphic dehydration reactions will occur at shallow depths beneath the forearc, limiting the amount of H<sub>2</sub>O available to flux melting in the mantle wedge beneath the arc. For these reasons, it could be that the degree of mantle melting is lower west of Adak compared to the central Aleutians. Could this explain why high Mg# andesites are abundant in the west?

High H<sub>2</sub>O and alkali contents stabilize high SiO<sub>2</sub> in olivine-saturated melts over a range of pressures [*Hirschmann et al., 1998; Kushiro, 1975; Ryerson, 1985*]. It might be inferred that low degrees of melting of mantle peridotite (perhaps with aqueous fluid present) could produce a melt which was highly enriched in H<sub>2</sub>O and alkalis, and thus would be andesitic rather than basaltic. This trend has been recently demonstrated in both nominally anhydrous and deliberately hydrous experiments [*Baker et al., 1995; Gaetani and Grove, 1998; Hirose, 1997; Kushiro, 1990*]. However, the maximum SiO<sub>2</sub> content attainable via this process is controversial [*Baker et al., 1996; Draper and Green, 1999; Falloon et al., 1996*].

In general, very small degree, alkali-rich melts of nominally *anhydrous* mantle lherzolite at pressures of 1 GPa and above tend to be nepheline normative, whereas primitive Aleutian

andesites are quartz normative. High H<sub>2</sub>O has a somewhat different effect from high alkali contents, and tends to raise the proportion of SiO<sub>2</sub> in mantle melts relative to NaAl and KAl components, yielding quartz-normative, primitive andesite liquids (calculated H<sub>2</sub>O-free, as we have done throughout this paper) at 1 to 1.5 GPa [*Gaetani and Grove, 1998; Hirose, 1997; Kushiro, 1990*]. However, by analogy with simple system phase equilibria (e.g., [*Kushiro, 1968; Kushiro, 1969*]), at pressures higher than 1.5 or 2 GPa it is expected that hydrous, small degree partial melts of lherzolite will be alkaline rather than quartz normative. Partial melts of harzburgite can be more SiO<sub>2</sub>-rich (and Al- and Ca-poor) than partial melts of lherzolite (e.g., [*Falloon and Green, 1987; Falloon and Green, 1988; Falloon et al., 1988; Walter, 1998*]). Thus, hydrous melting of harzburgite remains a plausible, relatively little-studied possibility for producing quartz-normative partial melts of mantle peridotite, even at pressures greater than 1.5 GPa. Two primitive andesite lava compositions, from Japan and California, have been experimentally shown to be in equilibrium with mantle harzburgite compositions at 1 to 1.5 GPa [*Baker et al., 1994; Tatsumi and Ishizaka, 1982*].

In summary, many of the major element characteristics of primitive Aleutian andesites could potentially arise as a result of small degrees of partial melting of spinel peridotites. However, the trace element contents of *enriched* primitive andesites do not seem to be a result of this process alone. If enriched, primitive andesites in the western Aleutians form by partial melting of spinel lherzolite similar to the mid-ocean ridge basalt source, or even by partial melting of more depleted, spinel harzburgite, then the enrichments of these magmas in Sr/Y, Dy/Yb and La/Yb must be explained via metasomatism of the mantle source prior to melting.

In general, based on results of thermal modeling, many petrologists have inferred that temperatures in subduction zones are too low to allow partial melting (see Section 4.8), and thus that transport of incompatible elements from subducted material into the mantle must be via an

aqueous fluid. This is the most prevalent explanation for all types of arc magmas, including enriched, primitive andesites [Grove *et al.*, 2001] but also including more common primitive basalts worldwide (e.g., [Gill, 1981]). A popular variant on this idea is that aqueous fluids carrying incompatible elements lower the solidus temperature in the mantle wedge and *cause* partial melting. In this view, the concentrations of trace elements in the fluid can be estimated from H<sub>2</sub>O contents of glasses combined with assumptions about the mantle source composition (e.g., [Class *et al.*, 2000; Eiler *et al.*, 2000; Grove *et al.*, 2001; Stolper and Newman, 1992; Miller *et al.*, 1992b]).

Strikingly, however, the fluid compositions inferred from this approach are very different from those observed in experiments on partitioning between aqueous fluids, mantle peridotite, and basaltic eclogite at 1 to 4 GPa (e.g., [Ayers, 1998]). As pointed out by Hawkesworth *et al.* several years ago, the concentrations of elements such as Th and the light REE in experimental aqueous fluids are one to three orders of magnitude lower than the concentrations that are required to explain arc magma enrichments [Hawkesworth *et al.*, 1993a; Hawkesworth *et al.*, 1993b]. Furthermore, experimental data predict very large fractionations of relatively soluble elements, such as Pb and Sr with fluid/rock ~100, from relatively insoluble elements such as Th and REE, with fluid/rock from ~ 0.01 to 1 [Ayers, 1998; Ayers *et al.*, 1997; Brenan *et al.*, 1996; Brenan *et al.*, 1995a; Brenan *et al.*, 1995b; Johnson and Plank, 1999; Kogiso *et al.*, 1997; Stalder *et al.*, 1998; Tatsumi *et al.*, 1986]. In contrast, inferred aqueous fluid compositions required to metasomatize the mantle source of enriched, primitive andesites show very little fractionation of Pb and Sr from light REE and Th [Grove *et al.*, 2001; Stolper and Newman, 1992]. Similarly, enriched, primitive andesites in the western Aleutians have correlated Pb and Th concentrations (Figure 22, and [Kay, 1980; Kay and Kay, 1994a; Yogodzinski *et al.*, 1994]). For reasons such as these, some investigators have suggested that enrichment of the arc mantle source in many incompatible elements, is via transport of a small degree partial melt of

subducted material (e.g., [Class *et al.*, 2000; Elliott *et al.*, 1997; Hawkesworth *et al.*, 1997; Plank and Langmuir, 1993]), and *not* via aqueous fluid metasomatism.

Keppler and co-workers (e.g., [Keppler, 1996]) found that NaCl-rich aqueous fluids can be “supercritical”, and therefore are completely miscible with H<sub>2</sub>O-rich silicate liquids at subduction zone pressures. If so, then the distinction between Cl-rich aqueous fluids and melts may be an artificial one. However, mechanical dewatering and metamorphic dehydration reactions, combined with compaction, are likely to remove much of the H<sub>2</sub>O in subducted material beneath the fore-arc region in most arcs. Since the solubility of NaCl in H<sub>2</sub>O generally increases with temperature, aqueous fluids hotter than seawater are likely to remove most of the NaCl from sediments and metabasalts at relatively shallow depths in the subduction zone. Therefore, it seems unlikely that NaCl-rich, supercritical fluids are added to the mantle source beneath volcanic arcs.

In summary, while it is likely that primitive andesites can be in major element equilibrium with shallow mantle peridotites, enriched, primitive andesites in the Aleutians probably were not produced by partial melting of spinel peridotite, preceded by aqueous fluid metasomatism of the mantle source. More likely, they were formed by reaction between silicate liquids – derived from deeper in the subduction system – and shallow mantle peridotite. During reaction at relatively high melt/rock ratios (~1), major elements would be modified to approach equilibrium with olivine + pyroxene + spinel, but the incompatible trace element pattern would be largely unchanged, and would reflect fractionation processes at greater depth.

#### **4.5 Residual garnet? Yes.**

The high Sr/Y, Dy/Yb and La/Yb ratios and low Y and Yb concentrations of enriched, primitive Aleutian andesites can be explained as the result of residual garnet in the source of melting (e.g., [Kay, 1978a; Yogodzinski *et al.*, 1995]). More specifically, high middle to heavy

REE ratios must have been produced by melting or crystal fractionation involving residual garnet. In melting, if the source had chondritic Tb/Yb and Dy/Yb ( $Tb/Yb(N) \approx Dy/Yb(N) \approx 1$ ), observed Tb/Yb(N) up to 4 and Dy/Yb(N) up to 2.5 in primitive Aleutian andesites require that the ratios of the bulk rock/liquid distribution coefficients,  $D(Tb/Yb)$  and  $D(Dy/Yb)$ , must be less than 0.25 and 0.4, respectively. This is only likely where the residual assemblage includes garnet.

The main phases controlling REE abundance in melting of mantle peridotite and basaltic eclogite compositions are garnet and clinopyroxene. Over the range of melt compositions from basalt to dacite, clinopyroxene/liquid  $D(Tb)$ ,  $D(Dy)$  and  $D(Yb)$  are all about  $\sim 0.3$  to  $0.8$ , with  $D(Tb/Yb)$  and  $D(Dy/Yb) \approx 1$  (e.g., data compiled at <http://www.earthref.org/>). Measured garnet/liquid  $D(Tb)$ ,  $D(Dy)$  and  $D(Yb)$  are about 1 to 20, 1 to 30, and 4 to 40, respectively, with  $D(Tb/Yb)$  from 0.1 to 0.75 and  $D(Dy/Yb)$  from 0.2 to 0.7 (<http://www.earthref.org/>). Within this range, the highest values of  $D(Tb)$ ,  $D(Dy)$ ,  $D(Yb)$ ,  $D(Tb/Yb)$  and  $D(Dy/Yb)$  are for partitioning between garnet and dacite. For example, Rapp and Shimizu [Rapp and Shimizu, 2001] determined garnet/liquid  $D(Dy)$  and  $D(Yb)$  from 1 to 22 and 6 to 40, respectively, with  $D(Dy/Yb)$  ranging from 0.36 to 0.55, in the products of experimental, hydrous partial melting of basaltic eclogites [Rapp *et al.*, 1999]. Bulk rock/liquid  $D(Tb/Yb)$  and  $D(Dy/Yb)$  of  $\sim 0.25$  and  $0.4$  can be attained with a garnet/clinopyroxene ratio in the source of  $\sim 1$  or more from a variety of combinations of the observed garnet/melt partitioning data. Basaltic eclogites and garnet-bearing mantle peridotites have a garnet/clinopyroxene ratio  $\sim 1$  or more (e.g., [Cox *et al.*, 1987; Rapp *et al.*, 1991] and F.R. Boyd, pers. comm. 1987).

We envision four different scenarios in which garnet plays an important role in the genesis of REE fractionation in enriched, primitive Aleutian andesites.

(1) Partial melting of unusually, hot, subducted basaltic rocks in eclogite facies gives rise to the enriched component in Aleutian primitive andesites (Figure 23B). The western Aleutian subduction zone may be unusually hot because of slow convergence rates (Section 1.5).

(2) In the unusually cold wedge in the western Aleutians, only deep-seated, garnet peridotites are hot enough to undergo fluid-fluxed partial melting (Figure 23A). The mantle wedge in the western Aleutians might be unusually cold and/or relatively H<sub>2</sub>O-poor because slow convergence leads to increased cooling of the wedge via conduction into the subducting plate, and because a hot subducting plate may undergo most of its dehydration at shallow depths, beneath the forearc area. Low temperatures would facilitate garnet stability to depths of ~ 50 km in the wedge. For example, basaltic melt was saturated in a garnet peridotite assemblage at 1.6 GPa and temperatures less than ~ 1250°C in a system with ~ 5 wt% H<sub>2</sub>O in the melt [*Gaetani and Grove, 1998*]. Higher H<sub>2</sub>O contents could lead to even lower pressures for garnet stability on the mantle peridotite solidus.

(3) Subducted sediments and/or basalts melt beneath the entire arc. In the eastern and central Aleutians this component is difficult to detect because the subduction zone melt component is swamped by abundant basalts produced by mantle melting. In the west, where the mantle is colder and fluxed by a smaller amount of aqueous fluid, basalts are rare and the subduction zone melt component is seen in its undiluted form.

(4) Cooling melts in the uppermost mantle might begin to undergo crystal fractionation while reacting with mantle peridotite (Figure 23C). Experimental data show that hydrous high Mg# andesite melts are garnet + pyroxene saturated at 1.5 GPa [*Carroll and Wyllie, 1989*]. If reaction between such melts and surrounding mantle peridotite produced a garnet pyroxenite while maintaining a high Fe/Mg ratio in derivative liquids, it could form a high Mg# light REE enriched, heavy REE depleted liquid product.

Crystal fractionation involving garnet in the lower crust cannot explain Aleutian primitive andesite compositions for several reasons. First, all of the most light- and middle-REE enriched andesites in the Aleutians are also the most primitive lavas, with Mg#'s greater than 0.6 and in some cases greater than 0.7. This seems to rule out an important role for crystal fractionation in the crust. Second, crustal thickness in the Aleutians is ~ 30 to 35 km [*Fliedner and Klemperer, 1999; Grow, 1973; Holbrook et al., 1999*]. Recent investigations on crystallization of a hydrous, primitive andesite composition show that such melts are not garnet saturated at 1.2 GPa [*Müntener et al., 2001*]. Garnet was stable in andesitic and basaltic bulk compositions only after large degrees of crystallization decreased the Mg# to less than 0.5. Thus, primitive melts probably are not garnet saturated at Aleutian Moho pressures.

The role of residual garnet seems clear, but it is difficult to determine from REE systematics whether residual garnet remains in subducted eclogite or in the mantle wedge. At the end of this section, we are left with four possible scenarios: (1) Partial melting of eclogite in a relatively hot subduction zone in the west. (2) Partial melting of garnet peridotite in an unusually cold mantle wedge in the west. (3) Partial melting of eclogite in the subduction zone throughout the Aleutians, obscured by extensive partial melting of relatively hot, H<sub>2</sub>O-rich mantle peridotite in the east. (4) Combined crystallization of garnet pyroxenite and reaction with peridotite in the uppermost mantle.

#### **4.6 Geodynamic setting of high Mg# andesites**

REE systematics discussed in the preceding section require a role for residual garnet in enriched, primitive andesite genesis, but do not allow us to determine whether the residual garnet resides in subducted crust or in the mantle wedge.

It might be expected that the amount of partial melting of subducted basalt would be larger in the Aleutians west of Adak, where the trench-orthogonal convergence rate is slow and

therefore subducted crust has a long time to heat conductively as it descends below the arc [Yogodzinski *et al.*, 1994; 1995]. Additionally, along the transcurrent plate boundary beneath the Komandorsky block the subducting plate may have a “torn”, exposed edge, and thus be heated by the convecting mantle from one side as well as from the top [Yogodzinski *et al.*, 2001].

However, while slow convergence and a “torn slab” in the western part of the arc may have produced an unusually hot subduction zone, this is probably combined with an unusually cold mantle wedge, and with a relatively low flux of H<sub>2</sub>O into the wedge beneath the arc. Which of these factors might be important in producing abundant, primitive andesites? In an attempt to resolve this, we turn to the geodynamic setting of the western Aleutians compared to other arcs worldwide.

Defant and Drummond identified a first-order correlation between the age of subducting crust and the nature of arc lavas worldwide [Defant and Drummond, 1990]. They found that Y and Yb concentrations in andesitic to dacitic lavas were relatively low where subducted crust was younger than ~ 25 Ma, and much higher where subducted crust was older. They interpreted this correlation to indicate that subducted basalt in eclogite facies undergoes partial melting when hot, young crust is subducted, but not when cold, old crust is subducted. It is important to note that Defant and Drummond incorrectly inferred that subducting crust beneath the Buldir area in the Aleutians was 15 Ma, whereas data show that in fact subducted crust beneath Buldir is more than 40 Ma. At the present time, the crust subducting beneath the Buldir area is ~ 45 to 50 Ma. When the dead Kula-Pacific spreading ridge slipped below the distal Aleutians, about 3 Myr ago, the thermal age of the slab was ~ 40 Myr [Atwater, 1989; Lonsdale, 1988]. Despite this discrepancy, if it could be shown that subduction of crust younger than, say, 45 Myr old led to formation of high Sr/Y, Dy/Yb, La/Yb magmas, while subduction of older crust did not, and that this was *independent of subduction rate*, then we could infer that subduction zone

temperature was the most important variable in the origin of enriched, high Mg# andesites, and thus that an eclogite melting component is essential in their genesis.

There are many localities with enriched, primitive andesites similar to those in the Aleutians: e.g., SW Japan, the Cascades, Baja California, Mexico, and the Austral Volcanic Zone in southern Chile [Defant *et al.*, 1991; Defant *et al.*, 1989; Hughes and Taylor, 1986; Luhr *et al.*, 1989; Puig *et al.*, 1984; Rogers *et al.*, 1985; Tatsumi and Ishizaka, 1982]. Unfortunately, in all of these other areas the subducting oceanic crust is younger than 25 Ma and trench-orthogonal convergence rates are less than 40 mm/yr. In a few areas (SW Japan, Mt. Shasta in the Cascades), primitive andesites clearly incorporate Pb derived from subducted sediment [Grove *et al.*, 2001; Shimoda *et al.*, 1998]. However, this could be a result of partial melting of subducted sediment alone, or of aqueous fluid transport of sediment Pb into garnet peridotites. Thus, it is difficult to separate the effect of subducting young crust – which could lead to a hot subduction zone and a relatively hot mantle – from the effect of slow convergence, which produces a hot subduction zone and a colder mantle.

There are areas in which young crust is subducting at high convergence rates, mainly along the Andean subduction zone to the north of the Austral Andes. With the exception of the Puna Plateau area [[Kay and Kay, 1994a], primitive andesites are not found north of the Austral volcanic zone. (The primitive andesites of the Puna Plateau have been interpreted as the result of delamination and partial melting of exceptionally thick lower crust [Kay and Kay, 1993]). However, because of the thick continental crust in the Andes, which may act as a “filter” preventing the rise of primitive magmas and/or modifying ascending melts, it is difficult to be sure whether the lack of primitive andesites in most of the Andes is significant. Perhaps, subduction of young oceanic crust alone is not sufficient to ensure the presence of abundant, enriched primitive andesite lavas.

The Lesser Antilles and the South Sandwich arcs have among the slowest convergence rates of any modern subduction zone [Argus and Gordon, 1991; DeMets et al., 1990; DeMets et al., 1994]. High Mg# andesites are found in neither arc. Oceanic crust older than 60 Ma is being subducted beneath the Lesser Antilles and the northern half of the South Sandwich arc. In addition, crust subducting beneath the southern half of the South Sandwich arc formed at the super-slow spreading American-Antarctic Ridge. Super slow spreading crust may incorporate very little basaltic material, being predominantly composed of serpentinites formed via alteration of mantle peridotite [Dick, 1989; Snow, 1995]. Also, super-slow spreading crust may be significantly colder, at a given age, than oceanic crust created at full spreading rates greater than  $\sim 0.01$  m/yr (e.g., [Henstock et al., 1993; Phipps Morgan and Chen, 1993; Reid and Jackson, 1981; Sleep, 1975]). In any case, the examples of the Lesser Antilles and the South Sandwich arc illustrate that slow convergence alone is insufficient to ensure the presence of abundant, enriched primitive andesites.

In summary, it appears that both slow convergence rates and young subducting crust may be necessary in order to produce abundant, erupted primitive andesites. Unfortunately, because these two variables cannot be separated, this is not helpful in determining whether residual garnet in primitive andesite genesis resides in the subducting crust or in the mantle wedge.

#### **4.7 Residual rutile? Yes, in partial melting of eclogite.**

Kay [1978] presented a simple model of a small degree of partial melting of MORB in eclogite facies, which accounted for the trace element characteristics of the highly enriched, high Mg# andesites on Adak. Noting that the Mg# and Ni contents of these lavas were much higher than in a small degree partial melt of MORB, Kay suggested that melts of subducted eclogite reacted with the mantle during transport. More recently, this model has been extended to include additional incompatible elements. In particular, rutile is residual during small to moderate

degrees of melting of MORB-like compositions in eclogite facies [*Rapp and Watson, 1995; Rapp et al., 1991*]. Elements such as Ta and Nb are compatible in rutile, and thus the marked depletion of Ta and Nb relative to La in enriched, high Mg# andesites is easily explained as the result of eclogite melting [*Kelemen et al., 1993; Ryerson and Watson, 1987; Yogodzinski et al., 1995*].

The stability of residual rutile during melting or crystal fractionation is dependent on many factors, including SiO<sub>2</sub> concentration in melts, temperature, and pressure as well as the bulk TiO<sub>2</sub> content of the system. It has been proposed that rutile might be stable in peridotite/melt systems (e.g., [*Bodinier et al., 1996*]). This might be possible under hydrous, low P conditions where melts might be hydrous, silica-rich, TiO<sub>2</sub>-rich, and relatively cold [[*Schiano and Clocchiatti, 1994; Schiano et al., 1995; Schiano et al., 1994*]. However, this is uncertain and has never been observed experimentally. In contrast, rutile saturation in low temperature, SiO<sub>2</sub>-rich, TiO<sub>2</sub>-rich melts of eclogite at ~ 3 GPa has been theoretically predicted and experimentally observed [*Bodinier et al., 1996; Rapp and Watson, 1995; Rapp et al., 1991; Ryerson and Watson, 1987*].

Using magmatic temperatures calculated as described in Section 2.1, we calculated TiO<sub>2</sub> concentration required for rutile saturation [*Ryerson and Watson, 1987*] in Aleutian primitive magma compositions, and compared it to observed TiO<sub>2</sub> concentration (Figure 24). At 1 GPa, with 0 to 10 wt% H<sub>2</sub>O in the magmas, 0 to 30% of 50 primitive Aleutian andesites had TiO<sub>2</sub> concentration within 1 wt% of the required concentration for rutile saturation. At 2 GPa with the same range of H<sub>2</sub>O, 0 to 40% of Aleutian primitive andesites had TiO<sub>2</sub> within 1 wt% of rutile saturation. Thus, in the pressure range in which primitive andesites could possibly be in major element equilibrium with mantle peridotite, their TiO<sub>2</sub> contents generally fall short of the value required for rutile saturation.

At higher pressure, as noted above, rutile becomes more stable. For example, at 3 GPa, 15 to 70% of the primitive andesites are within 1 wt% TiO<sub>2</sub> of rutile saturation. However, at these pressures, primitive andesites would be far from equilibrium with mantle peridotite.

To summarize, at low pressures, primitive andesites have too little TiO<sub>2</sub> to form rutile, while at high pressures the same melt compositions have too much SiO<sub>2</sub> to equilibrate with peridotite. Thus, it seems unlikely that residual rutile in the mantle is the cause of Ta and Nb depletions in primitive Aleutian andesites.

If primitive andesites are partial melts of eclogite which have been heated and modified during reaction with peridotite during ascent through the mantle wedge (Sections 2.6, 4.5, 4.7, 4.10, and 4.12), the initial melts prior to reaction would have been colder and more SiO<sub>2</sub>- and H<sub>2</sub>O-rich. Such melts, experimentally produced during partial melting of basaltic eclogites under H<sub>2</sub>O-undersaturated conditions, have been shown to be rutile-saturated [*Rapp and Watson, 1995; Rapp et al., 1991*]. Thus, we infer that a component in *all* primitive Aleutian andesites – regardless of the degree of light REE enrichment – was a rutile saturated partial melt of eclogite.

One could ask if there are different components in Aleutian primitive andesites, one carrying a Nb and Ta depletion, and another that is light REE enriched. However, despite the depletion of Ta and Nb relative to La in Aleutian magmas, Ta and Nb concentrations in primitive lavas increase from east to west, and both are weakly correlated with La concentration. Therefore it seems probable that Ta and Nb depletion form in the same manner as light REE enrichment, due to the combined presence of residual rutile and garnet in the source of the enriched component in Aleutian primitive andesites.

Nb and Ta are depleted relative to, e.g., La throughout the Aleutian arc, not just in primitive andesites. The causes of Nb and Ta depletion in arc basalts have been extensively debated. A variety of workers have reached the conclusion that the depletion is due to the presence of

residual rutile during partial melting of subducted sediment and/or basalt in eclogite facies (e.g., [Johnson and Plank, 1999; Kelemen et al., 1993; Ryerson and Watson, 1987]). If so, then most arc magmas must include a small component of eclogite melt, and Aleutian primitive andesites are simply distinguished by having a comparatively large proportion of such a component, as previously suggested by Kay [1980]. (We will return to this theme in Sections 4.10 and 4.11; also, please see Figure 23D, E and F).

Up to this point, we have reviewed a variety of possible explanations for genesis of enriched, primitive Aleutian andesites. At the end of Section 4.5, we were left with four possibilities in explaining the genesis of primitive Aleutian andesites, two of which involved partial melting of subducted eclogite. In this Section, we concluded that partial melting of subducted basalt in eclogite facies is required to explain depletions of Nb and Ta relative to REE in these lavas. Therefore, we can narrow the list to two possibilities : (1) Partial melting of eclogite in a relatively hot subduction zone in the west, and (2) partial melting of subducted eclogite throughout the Aleutians, obscured in the east by extensive partial melting of relatively hot, H<sub>2</sub>O-rich mantle peridotite. In order to discriminate between these two options, we must now consider the thermal state of subduction zones and the overlying mantle wedge, both for the Aleutians and on a worldwide basis.

## **4.8 Thermal models and partial melting in the Aleutian subduction zone**

### ***4.8.1 Extant thermal models predict low subduction zone temperatures for “normal” arcs***

Modern thermal calculations suggest that partial melting of subducted material is unlikely beneath the Aleutian arc. These calculations are essentially of three types: (1) analytical approximations including various assumptions about coupling between the subducted crust and the overlying mantle and about convection in the mantle wedge (e.g., [Molnar and England, 1990; Molnar and England, 1995]), (2) kinematic models, in which the thermal regime is

calculated numerically using analytical expressions for corner flow in the mantle wedge, with model results depending on various input parameters including the thickness of the arc “lithosphere” and the depth of coupling between subducting crust and overlying mantle (e.g., [Iwamori, 1997; Peacock, 1990a; Peacock, 1990b; Peacock, 1991; Peacock, 1996; Peacock and Hyndman, 1999; Peacock et al., 1994; Peacock and Wang, 1999; Pearce et al., 1992] and Peacock, this volume), and (3) dynamic models in which the mantle flow field as well as the thermal regime are calculated numerically, with model results depending on parameters such as mantle viscosity [Davies and Stevenson, 1992; Furukawa, 1993a; Furukawa, 1993b; Furukawa and Tatsumi, 1999; Kincaid and Sacks, 1997]. These models differ in many respects, but all agree that subduction of oceanic crust that is more than 20 Myr old at down-dip rates greater than 0.02 m/yr will not produce temperatures at the top of the subducting plate that are high enough to allow fluid-saturated melting of sediment or basalt. Fluid-undersaturated melting, at yet higher temperatures, is even less compatible with these thermal calculations.

#### ***4.8.2 Convergence rate and plate age “normal” beneath Aleutian volcanoes with “eclogite melt” component***

As noted above, young, enriched, high Mg# andesites (< 600,000 yrs old) are found in the vicinity of Buldir Island, particularly at the submarine volcano at 184.7°W from which dredge sample 70B29 was obtained. The down-dip subduction velocity beneath this site is estimated to be ~ 0.04 m/yr and the age of the plate beneath this site is estimated to be ~ 45 to 50 Ma. Thus, either the inference that enriched, high Mg# andesites contain a component formed by partial melting of subducted basalt is incorrect, there are special conditions in the subduction zone beneath Buldir (such as a “torn slab”; [Yogodzinski et al., 2001]), or all modern thermal calculations underestimate temperatures at the top of the subducting plate.

Additional evidence for a young, enriched high Mg# andesite component in the western Aleutians comes from plutonic xenoliths in Holocene lavas from Adak. These xenoliths include

clinopyroxene crystals with fine-scale oscillatory zoning, and interstitial glass that was quenched against the host lava [Conrad and Kay, 1984; Conrad *et al.*, 1983]. The preservation of fine scale zoning (zones with distinct Fe and Mg contents as small as 25 microns) and quenched interstitial glass both indicate that the xenoliths are not much older than their host lavas. Even assuming that the interstitial glasses were hydrous and therefore relatively low temperature, one can use a lower limit diffusivity of  $\sim 10^{-22}$  m<sup>2</sup>/s for Fe/Mg interdiffusion in clinopyroxene (e.g., summary in [Van Orman *et al.*, 2001]) to calculate that the 25 micron zones are less than  $\sim 200,000$  years old.

Our recent ion probe analyses of clinopyroxene in these xenoliths [Yogodzinski and Kelemen, 2000] show that some zones have Sr/Y as high and higher than the clinopyroxene phenocrysts in previously analyzed, enriched primitive andesites [Yogodzinski and Kelemen, 1998], suggesting that the xenoliths crystallized, in part, from Holocene, enriched primitive andesites. The down-dip subduction velocity beneath Adak is  $\sim 0.06$  m/yr and the age of the subducting plate beneath this site is  $\sim 45$  to  $50$  Ma. Tectonically, Adak overlies a normal Benioff zone (e.g., [Boyd and Creager, 1991; Boyd *et al.*, 1995]), not a “torn slab”. Thus, again, either enriched, high Mg# andesites do not contain a component formed by partial melting of subducted basalt, or modern thermal calculations underestimate temperatures at the top of the subducting plate.

#### ***4.8.3 Additional evidence for hot mantle wedge and subduction zones***

In the following paragraphs, we enumerate several other lines of evidence that suggest that, in fact, thermal calculations systematically underestimate temperatures, and overestimate mantle viscosities, beneath arcs.

##### ***4.8.3.1 Magma/mantle equilibration at high temperature and low pressure***

The conditions of equilibration between primitive arc lavas and mantle peridotite have been estimated by several investigators. This is particularly straightforward for basalts with known, low H<sub>2</sub>O-content. For such lavas, last equilibration with the mantle occurs at ~ 1300 to 1350°C and 1 to 1.5 GPa [Elkins Tanton *et al.*, 2001; Sisson and Bronto, 1998]. Similarly, Tatsumi *et al.* [Tatsumi *et al.*, 1983] made phase equilibrium experiments to determine the conditions at which primitive lavas from NE Japan could have equilibrated with mantle peridotite, over a range of possible H<sub>2</sub>O-contents. For some samples, they also inferred conditions of ~ 1300 to 1350°C and 1 to 1.5 GPa. Similar conclusions have also been reached in phase equilibrium studies of primitive, high aluminum basalts [Bartels *et al.*, 1991; Draper and Johnston, 1992].

There are no published thermal models for subduction zones which predict such high temperatures in the uppermost mantle, near the base of the arc crust. Instead, all published models predict temperatures less than 1300°C at depths less than 75 km (~ 2.5 GPa). Of course, the temperatures of magmatic equilibration with mantle peridotite do not have to lie along a steady-state geotherm. For example, melt in a transient magma chamber could equilibrate with peridotite wall rock that was heated to temperatures well above the ambient temperature at a given depth. Not all arc magmas have low H<sub>2</sub>O contents, and some could have equilibrated with the mantle at much lower temperatures (e.g., our Figure 3A). Nonetheless, if current estimates of H<sub>2</sub>O contents in primitive arc basalts are correct, the ubiquity of primitive basalts and basaltic andesites in arcs worldwide suggests that magmatic temperatures of 1200°C or more, at pressures less than ~ 1.5 GPa, must be common. In turn, this suggests that such basalts equilibrated with the mantle under steady-state conditions.

Note that advective heat transport via mantle diapirs might give rise to substantial variation in temperature at a given depth. While diapirs have been predicted in some dynamic models, their thermal consequences have not been determined. Furthermore, the mantle flow field beneath arcs may be inherently three-dimensional, whereas all thermal models for arcs to date

have been two-dimensional. This is an intriguing topic for future research. Diapirs rising as a result of thermal buoyancy might be hotter than the surrounding mantle, while diapirs rising as a result of melt-retention buoyancy - for melt generated at depth, e.g., at or immediately above the interface between subducted crust and overlying mantle - could be colder than the surrounding mantle.

#### *4.8.3.2 High heat flow over arcs, and steep heat flow gradient at arc front*

Heat flow measurements in arcs are uniformly high. Where they are available, two dimensional data arrays show that the high heat flow is present both beneath and between arc volcanic centers. Heat flow in fore-arcs is much lower, and the transition from high to low heat flow is generally along a sharp front parallel to the arc front [Blackwell et al., 1982; Honda and Uyeda, 1983]. Blackwell et al. inferred from such data that the temperature must be  $\sim 1000^{\circ}\text{C}$  at  $\sim 0.3$  GPa beneath the Oregon Cascades. If so, temperatures in the lower crust and uppermost mantle are likely to be even higher. Furthermore, the sharp transition in heat flow near the arc front, from  $40 \text{ mW/m}^2$  to  $10 \text{ mW/m}^2$  over 10 km ( $0.003 \text{ mW/m}^3$ ), seems to require an equally sharp transition from high temperatures ( $\sim 1200^{\circ}\text{C}$ ) maintained by advective heat transport in the uppermost mantle immediately beneath the arc to low temperature, static mantle (less than  $500^{\circ}\text{C}$ ) beneath the entire forearc. In contrast, most published thermal models do not include sharp thermal gradients. For example, the minimum distance between 1200 and  $500^{\circ}\text{C}$  isotherms in the thermal models of Peacock (this volume) is  $\sim 50$  km. Even the models of Furukawa [1993], who used a thermally dependent viscosity and a shallow depth for decoupling between mantle wedge and subducting plate in an attempt to fit the heat flow data, do not yield heat flow gradients of  $0.003 \text{ mW/m}^3$  near the volcanic front.

#### *4.8.3.3 Low seismic velocities in the uppermost mantle beneath arcs*

Teleseismic tomography for NE Japan reveals a 4 to 6% P-wave velocity anomaly in the uppermost mantle, immediately beneath the arc crust, which is continuous along the strike of the arc [Zhao and Hasegawa, 1993; Zhao and Hasegawa, 1994; Zhao *et al.*, 1992a; Zhao *et al.*, 1992b]. Similar results have been obtained for the Tonga arc [Zhao *et al.*, 1997]. While this could be an artifact due to undetected seismic anisotropy, preliminary analysis suggests that the effects of anisotropy are small (Zhao, pers. comm. 2000). 4 to 6 %  $V_p$  anomalies in the MELT region beneath the East Pacific Rise have been taken to indicate the presence of melt in mantle peridotite at temperatures greater than 1300°C [Forsyth *et al.*, 1998a; Forsyth *et al.*, 1998b]. A similar inference may be warranted for the mantle beneath NE Japan, though temperatures could be lower because  $H_2O$  will stabilize mantle melts at lower temperatures. The regionally extensive nature of the seismic anomaly suggests that it is a steady-state feature of the uppermost mantle beneath the arc.

#### *4.8.3.4 Lack of viscous coupling between subduction zone and arc*

Fourth, viscous coupling between the mantle wedge and the subducting plate should produce a Bouger gravity low over arcs [Zhong and Gurnis, 1992], whereas no systematic variation is observed in the Bouger gravity signal over arcs. This discrepancy can be resolved if the mantle wedge has a lower viscosity, permitting mechanical decoupling between the wedge and the subducting plate [Billen and Gurnis, ].

#### *4.8.4 Rapid thermal convection in a low viscosity mantle wedge?*

Thermal convection in the mantle wedge, at rates substantially exceeding the rate of subduction, can potentially resolve all these discrepancies between thermal models and observations. If the wedge viscosity is sufficiently low, cooling beneath the lithosphere and along the interface between the subducting plate and the overlying mantle could provide a

driving force for thermal convection. Convective velocities can be estimated using the following simple scaling rule based on boundary layer theory:

$$v = 0.271 \kappa / L [\rho g \alpha \Delta T L^3 / (\mu \kappa)]^{2/3}$$

([*Turcotte and Schubert*, 1982], equations 6-338 and 6-306) in which  $\kappa$  is thermal conductivity,  $10^{-6} \text{ m}^2/\text{s}$ ,  $L$  is a characteristic length scale for the shallow part of the mantle wedge,  $\sim 200 \text{ km}$ ,  $\rho$  is a reference density for the mantle of  $3300 \text{ kg/m}^3$ ,  $g$  is the gravitational constant,  $10 \text{ m/s}^2$ ,  $\alpha$  is the thermal expansion coefficient for mantle peridotite,  $2.4 \cdot 10^{-5} \text{ K}^{-1}$ ,  $\Delta T$  is the temperature contrast between the core of the mantle wedge and the subduction zone,  $\sim 1000 \text{ K}$ , and  $\mu$  is the dynamic viscosity in  $\text{Pa s}$ . Convective velocities might exceed subduction velocities of  $\sim 0.05 \text{ m/yr}$  if the viscosity in the wedge,  $\mu$ , were less than  $\sim 10^{20} \text{ Pa s}$ .

Convective velocities would be  $\sim 10$  times larger than subduction velocities if the viscosity in the wedge were  $\sim 5 \cdot 10^{18} \text{ Pa s}$ . Such viscosities are within the range measured for olivine that includes substantial dissolved hydrogen [*Hirth and Kohlstedt*, 1996; *Kohlstedt et al.*, 1996; *Mei and Kohlstedt*, 2000], as might be expected in the relatively  $\text{H}_2\text{O}$  rich mantle above a subduction zone, and for partially molten,  $\text{H}_2\text{O}$  -free peridotite with melt fractions greater than  $\sim 5\%$  [*Hirth and Kohlstedt*, 1995a; *Hirth and Kohlstedt*, 1995b; *Kelemen et al.*, 1997]. A combination of partial melt and dissolved  $\text{H}_2\text{O}$  would also produce mantle viscosities less than  $5 \cdot 10^{18} \text{ Pa s}$ . However, to date no dynamic model of a subduction zone has incorporated such low viscosities.

If thermal convection becomes faster than the plate convergence rate, thermal boundary layers in the mantle, beneath the crust and along the interface with the subducting plate, will become thinner (Figure 25). This, in turn, will lead to more rapid transfer of heat between the mantle, the crust and the subducting plate, and raise the mean, internal temperature of the wedge, lowering its mean viscosity. Both of these factors will promote even more vigorous convection.

As a result, it is expected that a thermally convecting wedge would have temperatures near 1300°C at the base of the crust, satisfying the mantle-magma equilibration, heat flow, and seismic constraints. A sharp edge between the convecting wedge and the static forearc mantle would give rise to the observed sharp drop in heat flow between arc and forearc. A lower viscosity wedge, due to the effect of H<sub>2</sub>O and higher mean temperatures in the wedge, would explain the absence of a consistent Bouger gravity anomaly over arcs.

Thermal convection in the mantle wedge that is more rapid than the subduction rate could also lead to much higher temperatures along the interface between the subducting plate and the overlying mantle. One can understand this by analogy to formation of hornfels along the margins of magma chambers. Where magma is static, the highest temperature which can be attained along magma chamber walls is the average of the initial wall rock and magmatic temperatures. However, where magma is convecting, the thickness of the conductive boundary layer along the walls is reduced, and wall rock temperatures approach the magmatic temperature. Thus, thermal convection of the wedge could raise temperatures at the top of the subducting plate well above the fluid-saturated solidus for basalt and sediment in eclogite facies.

In summary, currently available thermal models fail to account for a variety of important constraints on the temperature and viscosity structure of arcs. A clear alternative is that mantle viscosities are lower than has been assumed in dynamic models up 'til now, and that thermal convection in the wedge is more rapid than the subduction rate. If so, one result would be that temperatures in subducted metabasalt and metasediment would rise above the fluid-saturated solidus, permitting partial melting if fluid is available.

#### **4.9 Unusual, Miocene thermal conditions beneath the Aleutians? Probably not.**

As noted above (Section 3), limited data on Tertiary plutons in the Aleutians suggest that light REE enriched, high Mg# andesite compositions are more abundant in these plutons than in

Recent lavas. It is uncertain whether this is due to the limited number of plutons available for study, different emplacement depths for different magma types, or a temporal change in Aleutian magmatism.

Similarly, with one exception, the 18 Aleutian high Mg# andesites with the strongest light REE enrichment ( $\text{La/Yb} > 9$ ) are Miocene [*Kay et al.*, 1998; *Yogodzinski et al.*, 1995]. These samples, found on Adak Island and in the Komandorsky block, have  $\text{Sr/Y} > 50$  and other distinctive characteristics. They have been interpreted as partial melts of subducted basalt in eclogite facies, which underwent reaction with the mantle during transport to the surface [*Defant and Drummond*, 1990; *Kay*, 1978a; *Yogodzinski et al.*, 1995; *Yogodzinski and Kelemen*, 1998]. One might infer that partial melting of subducted basalt occurred during the Miocene but has since ceased.

However, the exception is a single dredged lava from a submarine cone near Buldir Island in the western Aleutians (sample 70B29, 184.7°W) has 63 wt%  $\text{SiO}_2$ , an Mg# of 0.56, 9960 ppm K, La/Yb of 9.8, and Sr/Y of 73 (or  $\text{Sr}/(9.6 \cdot \text{Yb})$  of 104). This sample was too young to date by conventional K/Ar methods [*Scholl et al.*, 1976]. In addition, as described in Section 4.8, clinopyroxenes in Holocene xenoliths in Adak lavas have the same, enriched trace element characteristics [*Yogodzinski and Kelemen*, 2000] as clinopyroxenes in enriched, primitive Aleutian andesites [*Yogodzinski and Kelemen*, 1998]. Thus, data from the distal Aleutians and the Adak area suggest that partial melting of subducted basalt in eclogite facies continues today.

It is unfortunate that such an important conclusion must be based largely on incomplete geochronological data from a single dredged sample. This reflects the poor knowledge of western Aleutian volcanism, and represents a worthy research goal for the future.

#### **4.10 Melting of subducted material throughout the arc, or only in the west? Throughout.**

Given that thermal models for subduction zones fail to account for many first-order observations, we cannot rely on them to discriminate between the possibility that (1) partial melting of eclogite occurs in a relatively hot subduction zone in the west, and the alternative that (2) partial melting of subducted eclogite occurs throughout the Aleutians, obscured in the east by extensive partial melting of relatively hot, H<sub>2</sub>O-rich mantle peridotite. Instead, we must return to arguments based on geochemistry in order to further constrain magma genesis in the Aleutians.

Recent work has suggested that partial melting of subducted sediment is a common process in the central Aleutians, the Marianas and the Lesser Antilles (e.g., [Class *et al.*, 2000; Elliott *et al.*, 1997; Hawkesworth *et al.*, 1997; Plank and Langmuir, 1993]). This hypothesis is based on the observation that some elements, including K and Th, are highly enriched in arc magmas with isotopic signatures indicative of a large sediment component in their source. Because Th is essentially immobile in aqueous fluids equilibrated with silicates, but is incompatible during partial melting of silicates, it seems that the sediment component in the arc source must be transported in a partial melt. (Also see Section 2.4).

Miller *et al.* used Ce/Pb versus <sup>207</sup>Pb/<sup>204</sup>Pb systematics in lavas from Umnak Island in the central Aleutians to argue that Pb derived from subducted basalts is transported into the mantle wedge in a low Ce/Pb fluid rather than a high Ce/Pb melt phase. This, combined with evidence for sediment melting described in the previous paragraph (and in Section 2.4) gave rise to the aphorism, “sediments melt, basalts dehydrate”. However, as we have shown in Section 2.5 and Figure 15, MORB-like Pb isotopes are observed in both high and low Ce/Pb Aleutian lavas. Because Ce/Pb is correlated with Dy/Yb and Sr/Y, thought to be indicative of an eclogite melt component in the Aleutians, we infer that the MORB-like Pb isotope signature in high Ce/Pb Aleutian lavas is derived from a partial melt of subducted basalt, not from the mantle wedge. While high Ce/Pb lavas are mainly found at and west of Adak, they are also present at Bogoslof and Amak Islands, behind the main arc trend in the central Aleutians.

Where sediments are present, their high Ba, Pb, Sr and REE contents and extreme isotope ratios (compared to MORB and the mantle) will dominate the trace element and isotopic characteristics of a partial melt derived from both sediment and basalt. Under these circumstances, it would be difficult or impossible to detect the basalt melt component. Only where sediments are rare or absent would it be possible to clearly detect a component derived from partial melting of basaltic eclogite.

At a given pressure, the fluid-saturated solidus should be at very similar temperatures in both basaltic and sedimentary bulk compositions, since both pelitic sediments and basalts in eclogite facies are composed of mixtures of garnet, omphacite, kyanite, coesite and phengite, plus various minor phases (Max Schmidt, pers. comm. 2000). This is borne out by experimental data [Carroll and Wyllie, 1989; Johnson and Plank, 1999; Lambert and Wyllie, 1972; Nichols *et al.*, 1994; Stern and Wyllie, 1973]. Thus, it is difficult to envision a realistic scenario in which subducted sediments consistently undergo partial melting while subducted basalts never do. Instead, it seems likely that both undergo partial melting, or both do not.

As noted by Gill [Gill, 1981], most arc magmas, including light REE enriched, calc-alkaline andesites, do not have REE slopes as steep as those in partial melts of eclogite. Gill inferred from this that partial melting of eclogite did not play an important role in arc petrogenesis. However, Kelemen *et al.* [Kelemen *et al.*, 1993] reasoned that reaction between ascending eclogite melts and spinel peridotite in the mantle wedge might modify REE patterns, particularly if the reaction involved increasing magma mass, as discussed in Section 4.12. Middle to heavy REE concentrations can be modified by such reactions to values for melt in equilibrium with spinel peridotite (Figure 23D, E and F). As a result of mass balance, light REE concentrations are modified much more slowly, and remain high. Thus, the light REE enriched, heavy REE depleted pattern characteristic of eclogite melts is transformed to produce a light REE enriched melt with a flat middle to heavy REE pattern, similar to common, light REE enriched arc

magmas, and to the estimated composition of continental crust. Such a pattern is typical for lavas throughout the Aleutians, and we infer that it is possible – though by no means certain – that the light REE enrichment in Aleutian magmas may be due to the presence of a cryptic eclogite melt component, modified by extensive reaction with spinel peridotite in the mantle wedge.

We find the incompatible trace element evidence for a partial melt component derived from subducted material in the central Aleutians persuasive [Class *et al.*, 2000]. Similarly, in Sections 4.5 and 4.7, we have argued that there is strong trace element evidence for partial melting of subducted eclogite in the western Aleutians. Therefore, we hypothesize that partial melting of subducted material is ubiquitous throughout the arc.

Subduction zone melts may react with mantle peridotite in ways that modify their trace element characteristics substantially. Perhaps, even arc tholeiites with unexceptional trace element patterns started out as partial melts of subducted basalt and sediment in eclogite facies. In this view, reaction between subduction zone melts and overlying peridotite may lead to large increases in magma mass, modifying and diluting the original melt composition (Figure 23F). Thus, an arc tholeiite might derive more than 90% of its K<sub>2</sub>O and light REE from a partial melt of subducted basalt, but 90% of its heavy REE and MgO from overlying, mantle peridotite.

Incidentally, we do not mean to imply that no aqueous fluid is derived from subducting material and incorporated into arc magmas. In the same arc tholeiite, perhaps 90% of the B is derived from a fluid. In this paper, we have not concentrated on characterizing an aqueous fluid component, except to note that some proposed fluid component indicators (e.g., high Ba/La) do not appear to be separable from proposed sediment melt component indicators (e.g., high Th/La) in the Aleutians (Section 2.5). This does not mean that there are no fluids in the Aleutian arc source, but it does make it much more difficult to discern their nature and mode of transport.

#### **4.11 Why are primitive andesites mainly observed in the west? Less mantle melting.**

If partial melting of subducted material is ubiquitous beneath the Aleutian arc, then why are the highly enriched trace element signatures indicative of an eclogite melt component most clearly observed in lavas at and west of Adak? One possible explanation is that the colder mantle wedge in the western Aleutians, convecting more slowly and fluxed by a smaller proportion of aqueous fluid, undergoes relatively little partial melting. As a result, in the western Aleutians partial melts of eclogite rise through and react with the surrounding mantle (Section 4.12), but generally do not mix with basaltic magmas derived from decompression melting of peridotite. Further east, the hotter mantle wedge undergoes extensive melting, due to rapid convective upwelling and abundant hydrous fluids derived from the colder subducting plate. In this region, the mantle-derived basalts are far more voluminous than the component derived from partial melting of subducted material. As a result, the subduction zone melt component is difficult to detect unambiguously, except via enrichments of highly incompatible and insoluble elements such as Th.

Note that, in saying that the mantle wedge above a hot subduction zone would be “less hydrous”, we do not mean to imply that the melts produced in such an environment would be anhydrous. It is very unlikely that partial melting of subducted sediments or oceanic crust would occur in the absence of substantial quantities of H<sub>2</sub>O, even in “hot” subduction zones. Sources of H<sub>2</sub>O for hydrous partial melting could include (1) dehydration of talc and/or serpentine in peridotite underlying subducted crust (e.g., [Peacock *et al.*, 1994]), and/or (2) aqueous fluid trapped in pore space in eclogites. With regard to (2), experiments show that under hydrostatic conditions H<sub>2</sub>O does not form an interconnected grain boundary network in clinopyroxene-rich rocks until the porosity exceeds 7 volume % [Watson and Lupulescu, 1993]. We believe that it is most likely that subduction zone anatexis occurs under aqueous fluid saturated conditions, and is limited to small extents of partial melting (less than 10%).

#### 4.12 Reaction between partial melts of subducted eclogite and overlying mantle peridotite

Small degree partial melts of eclogite are far from equilibrium with mantle olivine. At constant temperature and pressure, siliceous melts of eclogite will react with olivine to produce pyroxene ( $\pm$  garnet) and modified, lower SiO<sub>2</sub> melt. In simple chemical systems, and probably in some natural systems, such reactions consume more liquid than they produce, ultimately leading to complete solidification, or “thermal death” (Figure 26) [Rapp *et al.*, 1999; Yaxley and Green, 1997]. It is possible that some eclogite melts do undergo thermal death, and then hybridized, pyroxene-rich mantle peridotite circulates to some other part of the mantle wedge, later melting to give rise to an enriched magma (e.g., [Ringwood, 1974; Yagodzinski *et al.*, 1994]). However, this scenario seems quite complicated, and therefore unlikely to be the general process of primitive andesite genesis.

Instead, we prefer a model in which the eclogite melt component interacts with mantle peridotite in the wedge, and is modified but never completely solidified. The observation of <sup>230</sup>Th excess in enriched, primitive andesites from Mt. Shasta in the Cascades [Newman *et al.*, 1986] and from lavas in the Austral Volcanic Zone in southernmost Chile [Sigmarsson *et al.*, 1998] provides an important constraint on the transport time of eclogite melt from source to surface. Radiogenic <sup>230</sup>Th excess is produced by small degrees of partial melting under conditions in which daughter <sup>230</sup>Th is less compatible in residual solids than parent <sup>238</sup>U. Present knowledge indicates that U is more compatible than Th in garnet [Beattie, 1993; Hauri *et al.*, 1994; LaTourrette and Burnett, 1992; Salters and Longhi, 1999] and - perhaps - high pressure, Na-rich clinopyroxene [Wood *et al.*, 1999], while in other mantle and eclogite phases Th is more compatible than U. Thus, the <sup>230</sup>Th excess in enriched, primitive andesites is consistent with derivation of the Th enriched component in such magmas by partial melting of subducted eclogite. The half life of <sup>230</sup>Th is  $\sim$  75,000 yrs. After a few half lives, excess <sup>230</sup>Th will decay, and <sup>230</sup>Th/<sup>238</sup>U ratios will return to steady-state, “secular equilibrium”. Therefore,

one can infer that the eclogite melt component in primitive andesites from Mt. Shasta and the Austral Volcanic Zone was transported from the subduction zone to the surface in less than ~ 300,000 years. This is plausible for magma transport via porous flow, hydrofracture, or, perhaps, in partially molten diapirs, but implausible for transport via solidification, solid state convection to a hotter part of the mantle wedge, and remelting.

As emphasized by Kelemen and co-workers [Kelemen, 1986; Kelemen, 1990; Kelemen, 1995; Kelemen *et al.*, 1993], a key to considering reaction between eclogite melt and overlying peridotite is that above subduction zones, melts must heat up as they rise and decompress (Figure 27, and [Kelemen and Hirth, 1998]). Grove *et al.* [Grove *et al.*, 2001] have recently made a similar point for partial melts of mantle peridotite deep within the mantle wedge. We first consider the effect of decompression heating on silicate liquid which is not reacting with solid rock, for example for ascent of liquid in a large fracture. The combined effect of increasing temperature and decreasing pressure will move the melt away from its liquidus. This is often referred to as “superheating”. (Although “superheat” in chemistry refers to a metastable phase above its thermal stability, for example liquid H<sub>2</sub>O at a temperature above its boiling point, we will follow convention in igneous petrology and use the term “superheat” in this paper to refer to a melt that is heated above its liquidus). Superheated melts would be able to dissolve a considerable mass of peridotite. Adiabatic ascent of silicate liquid creates a melt with ~ 65°C of superheat per GPa of pressure decrease, given an adiabatic gradient of ~ 10°C/GPa [Yoder, 1976] and a liquidus slope of ~ 75 °C/GPa for typical, mafic silicate melts [Morse, 1980]. Heating the same magma over the same pressure interval moves it still farther from its liquidus.

Equilibration of rising melt with surrounding rock is equivalent to a stepwise process in which melt rises through some finite interval of decompression without reacting, and then superheated melt reacts with surrounding rock at constant pressure. Reaction between superheated melt and rock will dissolve rock until cooling and the change in composition bring

the modified melt to its liquidus surface. Cooling of reacting melt will occur due to heat conduction from melt to surrounding rock and as a result of the heat of fusion .

The heat of fusion for mantle peridotite may be estimated to be ~ 150 cal/gm (e.g., [*Ghiorso and Carmichael*, 1980]), and the heat capacity of silicate melt is ~ 0.3 cal/gm°C (e.g., [*Stebbins et al.*, 1984]). As a result, dissolving 1 gm of rock in 100 gm of melt would cool the system about 5°C. However, as melt dissolves peridotite, the melt will cool *and* the liquidus temperature of the modified melt will increase, so that more than 5°C of superheat is required to dissolve 1% peridotite in silicate melt. Therefore, instead, we will assume that 10°C of superheat is required to dissolve 1% peridotite. With this assumption, adiabatically ascending melt could dissolve ~ 1% peridotite per 0.1 GPa of decompression, or ~ 20% peridotite over 60 km of ascent. Heating of ascending melt from ~ 800°C, just above the solidus of subducted material in the subduction zone, to ~ 1300°C in the hottest part of the mantle wedge, could supply energy for dissolution of an additional 10 to 100% peridotite, depending on the change in the liquidus temperature of the modified melt. Thus, one can see that an ascending, heating partial melt of subducted material, heating and reacting with mantle peridotite, could increase in mass and be dramatically modified in composition. If the surrounding mantle peridotite is also rising and undergoing decompression melting, mixing of modified subduction zone melts with “pure” mantle melts, and reaction of these mixtures with peridotite, is also likely.

Once reacting, cooling melts become saturated in solid phases, continued reaction at constant temperature, or in adiabatic conditions, may involve no change in liquid mass (“zone refining”), an increase in liquid mass if the reactions are of the form liquid 1 + pyroxene = liquid 2 + olivine [*Daines and Kohlstedt*, 1994; *Kelemen*, 1990], or a decrease in liquid mass if reactions are of the form liquid 1 + olivine = liquid 2 + pyroxene. This is because the heat of fusion of pyroxenes, per gram, is less than the heat of fusion of olivine per gram [*Ghiorso and Carmichael*, 1980]. (Where modified melts become saturated with olivine, reaction of the form

liquid 1 + pyroxene = liquid 2 + olivine will continue until modified melts are also saturated in pyroxene). Because reactions of the form liquid 1 + olivine = liquid 2 + pyroxene consume magma, a slowly ascending, pyroxene-saturated melt reacting with mantle peridotite could undergo “thermal death” [Rapp *et al.*, 1999; Yaxley and Green, 1997].

In the second to last paragraph, we described cases in which magma mass could increase during reaction with the mantle wedge above a subduction zone. Conversely, in the previous paragraph, we have described cases in which magma mass would decrease. Clearly, the actual outcome depends on the thermal regime (Section 4.8) and the mode of melt transport in the mantle wedge (Section 4.13).

In considering the process of thermal death, a further complication is introduced by the effect of alkalis and H<sub>2</sub>O on silicate liquidus surfaces. As magma mass decreases, incompatible element concentrations in the remaining melt increase. Where magma is saturated in amphibole and/or micas, alkalis and H<sub>2</sub>O may be compatible elements, and one can imagine that incompatible trace elements and CO<sub>2</sub> finally escape in an immiscible fluid as melt solidifies. However, at temperatures above the thermal stability limit of amphibole and mica (~ 1100°C; e.g., [Holloway, 1973]), alkalis and H<sub>2</sub>O are highly incompatible, and will become concentrated in remaining melt during the initial stages of thermal death. Because high alkali and H<sub>2</sub>O contents lower the liquidus temperature of silicate systems, this will tend to stabilize the presence of a melt within the mixed mantle peridotite + liquid system (Figure 28). Furthermore, because high alkali and H<sub>2</sub>O contents stabilize olivine relative to pyroxenes [Hirschmann *et al.*, 1998; Kushiro, 1975; Ryerson, 1985], as the modified liquid becomes enriched in these components it will move closer to equilibrium with mantle peridotite. Of course, if the melt reacts with sufficient peridotite, even alkalis and H<sub>2</sub>O can be dissolved in nominally anhydrous phases such as olivine, pyroxenes and garnet. However, this requires melt/rock ratios less than 0.01. Thus, if

rising subduction zone melts can reach temperatures of 1100°C or more without solidifying, it seems that they could survive and continue to ascend even if melt transport was slow.

#### **4.13 Transport of melts of subducted material through the mantle wedge**

Dissolution rates of mantle minerals in superheated, advecting melt have been experimentally measured, and range from  $10^{-8}$  to  $10^{-10}$  m/s [Brearley and Scarfe, 1986; Kuo and Kirkpatrick, 1985a; Kuo and Kirkpatrick, 1985b; Zhang *et al.*, 1989]. Magma with a viscosity less than 1000 Pa s, rising in an 0.1 to 1 m fracture at rates greater than  $\sim 0.01$  to 10 m/s [Turcotte and Schubert, 1982], would have little time to dissolve or diffusively equilibrate with surrounding peridotite. Since the primitive andesite magmas which carry the trace element signature of eclogite melting show signs of having equilibrated with mantle peridotite - at least in terms of Fe/Mg and Ni exchange - we infer that they did not ascend in cracks all the way from the subduction zone to the crust. However, it is possible to envision scenarios in which melt is initially transported in cracks that terminate within the mantle - e.g., in the hottest part of the mantle wedge, or in reservoirs just below the base of the crust - and then react with surrounding peridotite during slower transport by porous flow, in diapirs, or during residence in a sub-crustal magma chamber. Unfortunately, while such scenarios are easy to imagine, they are not readily amenable to further, quantitative evaluation.

Grove *et al.* [Grove *et al.*, 2001] considered the transport of hydrous, low density “basaltic” melts ( $\sim 2200$  kg/m<sup>3</sup>; [Ochs and Lange, 1999]) with a viscosity of  $\sim 10$  Pa s through peridotite via porous, grain boundary flow using the permeability estimates of Von Bargen and Waff [Von Bargen and Waff, 1986]. They also quantified likely ascent rates via Stokes flow for partially molten diapirs with a large density contrast (1100 kg/m<sup>3</sup>) with the surrounding mantle, for a hydrous mantle viscosity of  $10^{18}$  Pa s [Hirth and Kohlstedt, 1996]. They concluded that ascent rates greater than 1 m/yr are likely for porous flow with porosities greater than  $\sim 1\%$ , or for

diapirs with radii greater than ~ 3 km. Recall that ascent velocities greater than ~ 1 m/yr are required to preserve  $^{230}\text{Th}$  excesses observed in some primitive andesites, if  $^{230}\text{Th}$  excesses are formed during partial melting of eclogite at ~ 100 km depth.

In grain boundary porous flow, melts moving at velocities of ~ 1 m/yr in rocks with 1 mm to 1 cm grain size are likely to equilibrate with the surrounding rocks on the grain scale (e.g., Table 1 in [Aharonov *et al.*, 1995]). Viscosities of silica-rich partial melts of subducted sediment or basalt (> 60 wt%  $\text{SiO}_2$ ) could be more than two orders of magnitude higher than for basaltic melts but, conversely, high  $\text{H}_2\text{O}$  contents in silica-rich melts could yield viscosities similar to  $\text{H}_2\text{O}$ -free basalts [Shaw, 1972]. It is unlikely that melt viscosities would be substantially less than 1 Pa s. Even with porosities of 10%, viscosities of 1 Pa s, and a mantle grain size of 1 cm, melt velocities will not exceed ~ 2500 m/yr. If melt is homogeneously distributed along olivine grain boundaries, thermal death due to reaction of pyroxene + amphibole or mica saturated liquid with surrounding peridotite at temperatures less than 1100°C can probably occur in a few days (Section 4.12). Thus, if there is a boundary layer a few hundred meters thick in the mantle wedge above a subduction zone that is cooler than ~ 1100 °C, and melts of subducted material migrate entirely via porous flow along grain boundaries, thermal death is quite likely.

Ascent of partially molten diapirs is often invoked to explain melt transport beneath arcs. For example, Marsh, Myers and their co-workers proposed that diapirs of partially molten, subducted basalts rise through the mantle wedge (e.g., [Brophy and Marsh, 1986; Marsh, 1982b; Marsh and Kantha, 1978; Myers *et al.*, 1985; Singer *et al.*, 1989]). To overcome the high density of basalt in eclogite facies, they proposed that the diapirs were more than 20% molten at the beginning of ascent. Based on our calculations, dense eclogite solid assemblages would have to include more than 50% low density,  $\text{H}_2\text{O}$ -rich silicate melt in order to rise in 10 km wide diapirs at rates greater than 1 m/yr. To produce 20 to 50% melting of basalt at ~ 3 GPa requires temperatures greater than 1100°C, even in  $\text{H}_2\text{O}$ -saturated conditions (e.g., [Yoder, 1976]). To

produce a 10 km “sphere” of basalt with such a high melt fraction would require temperatures greater than 1100°C extending through the entire subducted crustal section. This is dramatically at odds with thermal models for subduction zones. While we believe that published thermal models are fundamentally flawed, and that their prediction of subduction zone temperatures are probably too low (see Section 4.8), it seems premature to imagine that subduction zone temperatures in excess of 1100°C are attained at the base of subducted crust beneath the Aleutians. (For example, for subducted crust beneath the Aleutians, Peacock (this volume) predicts ~ 500°C at the top of the subducted crust and ~ 300°C at the base). Thus, we infer that diapiric ascent of partially molten eclogite has not occurred beneath the Aleutians. With that said, we believe that diapiric ascent of *melts* derived from subducted eclogite, and perhaps of low density, subducted sediments in the solid state, could occur. Thus, for example, the work of Singer et al. [Singer et al., 1989] on ascent of eclogite melts through conduits composed of hybrid reaction products is very relevant in the context of this paper.

Particularly pertinent to this discussion is the scenario proposed by Ringwood [Ringwood, 1974] in which partial melts of subducted material rise into the mantle wedge, reacting with mantle peridotite, and then form diapirs of modified solid (pyroxenite) + melt (Figure 29). These rise, and the solids partially melt as they decompress and heat within the mantle wedge. The local melt/rock ratio in an incipient diapir could become high near the interface between mantle and subducted material, forming a hybrid pyroxenite and preserving the melt from thermal death. Further, prior to initiation of diapiric ascent, melt could accumulate in a pyroxenite reaction zone, perhaps beneath a permeability barrier formed by crystallization during melt/rock reaction (e.g., [Kelemen and Aharonov, 1998], especially their Figure 2). Thus, the melt fraction in a hybrid, pyroxenite reaction zone could become very high as a result of slow accumulation of small degree melts, rather than via large degrees of partial melting of subducted basalt and/or sediment.

Given that pyroxenite densities are similar to peridotite densities, the buoyancy force driving such diapirs must come from the presence of partial melt. To achieve ascent velocities of  $\sim 1$  m/yr via Stokes flow through H<sub>2</sub>O-rich mantle with a viscosity of  $\sim 10^{18}$  Pa s, given a melt with a density of  $\sim 2200$  kg/m<sup>3</sup>, pyroxenite diapirs with a diameter of 10 km would have to be  $\sim 30\%$  molten (diapirs with a diameter of 6 km would have to be  $\sim 100\%$  molten). At such high porosities, porous flow of melt within diapirs, and out into the surrounding rock, could occur. If pyroxenite diapirs “leaked”, they would lose the driving force for ascent. However, if the melt viscosities were very high, or if the diapirs were surrounded by a low permeability reaction zone, then this might not occur. We leave it to the reader to determine whether such scenarios are reasonable.

As pointed out by Grove et al. [*Grove et al.*, 2001], diapirs with diameters of 6 to 10 km would not thermally equilibrate with the surrounding mantle. Grove et al. sought to explain the high temperatures of many subduction zone magmas with a single model of melt transport. Because some arc melts record mantle equilibration temperatures in excess of 1300°C, and temperatures near the interface between mantle and subducted material may be lower than, e.g., 900°C, Grove et al. found that diapirs were an unlikely mode of melt transport from deep in the mantle wedge to the base of the crust. However, in this paper we focus on primitive andesites which record melt/mantle equilibration temperatures of 1100°C or lower.

An odd aspect of low temperature, primitive andesites is that they are found in relatively close proximity to other, basaltic magmas which have much higher mantle equilibration temperatures (e.g., [*Elkins Tanton et al.*, 2001]). Advective transport of deep melts in diapirs offers a possible explanation for this observation. Note that the chemical density contrast between diapir and mantle,  $\sim 100$  to  $1000$  kg/m<sup>3</sup>, is much larger than the density contrast due to a  $\sim 200$  to  $500^\circ\text{C}$  temperature difference,  $\sim 15$  to  $40$  kg/m<sup>3</sup>. As a consequence, a cold, chemically

buoyant diapir can rise through hot mantle peridotite. (In fact, the low mantle viscosity of  $10^{18}$  Pa s used in these calculations requires high mantle temperatures as well as high H<sub>2</sub>O contents).

Thus, in one scenario, higher temperature, basaltic magmas are derived from the surrounding mantle wedge at ~ 1300°C, and primitive andesites are transported in colder, chemically buoyant diapirs. As these diapirs ascend, they will heat – though not to ambient mantle temperatures – and decompress, undergoing extensive partial melting. At some point, some melts may escape and equilibrate with mantle peridotite rather than with hybrid pyroxenite. However, other melts may remain within high Mg# pyroxenite until they reach the base of the crust.

An alternative explanation for the different mantle equilibration temperatures for primitive basalts and andesites at nearby volcanic centers would be that the basalts are transported rapidly from the hot part of the mantle wedge into the crust, whereas the andesites ascend more slowly and equilibrate with the uppermost mantle. While this is an appealing hypothesis, mantle equilibration pressures for arc basalts are commonly close to the pressures at the base of the arc crust [Bartels *et al.*, 1991; Draper and Johnston, 1992; Elkins Tanton *et al.*, 2001; Sisson and Bronto, 1998; Tatsumi *et al.*, 1983]. There is little room for equilibration with *shallower* mantle than this. Thus, odd as it may seem, we prefer scenarios in which primitive andesites undergo rapid transport from a deep, cold source – in cracks or diapirs – passing through the hottest part of the mantle wedge without complete thermal equilibration.

To summarize, transport of subduction zone melts in cracks, followed by slow transport and reaction of these melts with shallow mantle peridotite somewhere in the wedge, offers one possible scenario for producing primitive andesites with an eclogite melting signature. Alternatively, subduction zone melts may form hybrid, pyroxenite reaction zones in the deepest part of the mantle wedge, and then these may rise through the wedge as partially molten diapirs. Either of these transport mechanisms might be fast enough to avoid thermal equilibration

between subduction zone melts and the hottest part of the mantle wedge. As a consequence, such transport processes could explain the relatively low melt/mantle equilibration temperature of primitive andesites as compared to nearby, primitive basalts in the same arc. Both transport mechanisms would allow rising eclogite melts to avoid thermal death, and ultimately to attain or closely approach chemical equilibrium with mantle olivine.

## **5. IMPLICATIONS FOR CONTINENTAL GENESIS**

### **5.1 Reaction between eclogite melt and mantle peridotite**

Figures 23C and D illustrate results of trace element models of reaction between a small degree partial melt of subducted eclogite and spinel peridotite under conditions of increasing melt mass. As previously proposed by [Kelemen *et al.*, 1993], the trace element contents of calculated liquids are strikingly similar to the estimated composition of the continental crust. Similarly, major element contents of liquids produced by experimental reaction of silicic melt with mantle peridotite closely approach those of high Mg# andesites and continental crust [Carroll and Wyllie, 1989], and of primitive andesites [Rapp *et al.*, 1999]. Thus, we believe that reaction of melts of subducted eclogite with the mantle wedge is a viable model for continental genesis.

In a parallel line of reasoning, many investigators have suggested that unusually orthopyroxene-rich mantle xenoliths derived from Archean cratons underwent SiO<sub>2</sub> enrichment as a result of reaction between highly depleted mantle residues and rising partial melts of subducted eclogite [Kelemen *et al.*, 1993; Kesson and Ringwood, 1989; Rudnick *et al.*, 1994]. Recently, Kelemen *et al.* [Kelemen *et al.*, 1998] showed that this hypothesis was compatible with the observed correlation of Ni in olivine with modal orthopyroxene in cratonic mantle xenoliths worldwide, and emphasized that orthopyroxene-rich mantle xenoliths cannot be simple residues of high pressure partial melting of primitive mantle peridotite. Thus, both continental crust and

cratonic mantle peridotites may be related by a process of reaction between subduction zone melts and sub-arc mantle peridotites.

A point of clarification is required here. Enriched primitive andesites (e.g., [Kay, 1978a]), partial melts of eclogite (e.g., [Gill, 1981] and our Figure 23B), and products of small amounts of reaction between eclogite melts and mantle peridotite under conditions of decreasing magma mass [Rapp *et al.*, 1999], all have middle to heavy REE slopes that are much steeper than those in typical Aleutian, calc-alkaline andesite (Figure 6), and much steeper than those estimated for continental crust (Figure 6A). Thus, in detail, we disagree with numerous authors who have proposed that direct partial melts of eclogite – without magma/mantle interaction – form an important component in the continental crust (e.g., [Defant and Kepezhinskis, 2001; Drummond and Defant, 1990; Martin, 1986; Martin, 1999; Rapp and Watson, 1995; Rapp *et al.*, 1991]). We believe that enriched, primitive andesites have undergone relatively little reaction with surrounding mantle. More extensive reaction, with gradually increasing melt mass and melt/rock ratios  $\sim 0.1$  to  $\sim 0.01$ , is required to increase heavy REE concentrations to the levels observed in most calc-alkaline andesites and in continental crust.

At small melt/rock ratios, such as are required to explain REE systematics, compatible element concentrations and ratios (e.g., Mg#, Ni) would be completely modified to values in equilibrium with mantle peridotite, and thus the liquids parental to high Mg# andesites and continental crust are inferred to be primitive andesites with  $\text{Mg\#} > 0.6$ . Thus, in addition to the reaction of eclogite melt with mantle peridotite, an additional process is required to explain the relatively low Mg# of continental crust (0.45 to 0.55) compared to primitive andesite. We believe the most likely explanation is that primitive andesites fractionate ultramafic cumulates in the uppermost mantle or at the base of the crust [DeBari and Coleman, 1989; Miller and Christensen, 1994; Müntener *et al.*, 2001], followed by “delamination” and return of these

cumulates to the convecting mantle [Arndt and Goldstein, 1989; Jull and Kelemen, 2001; Kay and Kay, 1991; Kay and Kay, 1993; Tatsumi, 2000].

## **5.2 Archean versus present-day thermal regime beneath arcs**

It has often been proposed that Archean continental crust was formed by a process similar to that which forms “adakite” magmas today. Martin [Martin, 1986] suggested that strong light REE enrichment and heavy REE depletion in Archean granitoids was the result of ubiquitous partial melting of eclogite in hotter, Archean subduction zones. Archean subduction zones may have been hotter than today’s because of higher mantle potential temperatures and faster rates of mantle convection (and plate tectonics) in the past. In support of Martin’s hypothesis, Defant and Drummond [Defant and Drummond, 1990] noted that modern “adakites” and arc granitoids, both with trace element contents similar to Archean granitoids, are found mainly in arcs where young oceanic crust (< 25 Ma) is being subducted. They inferred that this is due to partial melting of eclogite in unusually hot, modern subduction zones.

In this paper, we have developed the idea that partial melting of subducted sediment and/or basalt may be common in modern subduction zones. This is supported by a variety of trace element evidence from arcs worldwide. We argue that what is uncommon at present is a relative lack of basaltic magmas generated by partial melting of mantle peridotites in arcs such as the western Aleutians, the Cascades, parts of Central America, the Austral volcanic zone in Chile, and SW Japan. We suggest that it is the lack of basaltic melts in a few unusual localities, due to cooling of the mantle wedge and extensive shallow dehydration of the subducting plate at slow convergence rates, which makes it possible to detect end-member magmas with a trace element signature reflecting a large input from partially molten eclogite.

If primitive andesites in modern arcs are the result of partial melting of subducted eclogite and a *cold* mantle wedge, why should this form magmas analogous to those which played a large

role in *hot*, Archean crustal genesis? Following Martin [Martin, 1986] and many subsequent workers, we believe that Archean subduction zones were hotter than today, on average, due to higher mantle potential temperatures and more rapid rates of plate formation and subduction. Thus, most dehydration reactions in subducting volcanic rocks and sediments would have occurred at shallower depths than today, reducing the amount of H<sub>2</sub>O available to flux partial melting in the mantle wedge. In addition, we believe that the Archean shallow mantle was much more depleted in basaltic components than at present. For example, there is good evidence that Archean decompression melting, beneath ridges and/or hotspots, extended to 40% melting at shallow depths, and produced residues with Mg#’s of 0.93 or more (e.g., [Bernstein *et al.*, 1998]). Such refractory residues can undergo very little additional melting, even under hydrous conditions. For these reasons, there may have been very little melting in an Archean mantle wedge even though mantle potential temperatures in the Earth’s interior may have been much higher than at present. Thus, partial melts of subducted volcanic rocks, reacting with overlying mantle peridotite, might have constituted a large proportion of the total magma flux in Archean arcs.

## 6. CONCLUSIONS

High Mg# andesites are abundant at and west of Adak at ~ 174°W. In this region, there is a systematic along-strike increase in SiO<sub>2</sub> from east to west. The Aleutian high Mg# andesite lavas are important because they are similar in major and trace element composition to the continental crust. Such lavas are rare or absent in intra-oceanic island arcs other than the Aleutians. Also, the western Aleutians show the smallest influence of a subducted sediment component of any part of the Aleutians, so that it is unlikely that enrichments in elements such as U, Th, K and light rare earths are due to recycling of subducted, continental sediments. Thus, it appears that juvenile continental crust is being produced in the western Aleutians.

As proposed by Kay [Kay, 1978a], enriched, primitive andesites in the western Aleutians are produced by partial melting of subducted oceanic crust in eclogite facies, followed by reaction of these melts with peridotite in the overlying mantle wedge. Such melts are probably formed beneath the central Aleutians as well as the western Aleutians. In the central Aleutians, they are obscured by coeval partial melts of subducted sediment, with higher incompatible element contents, and by mixing with abundant tholeiitic basalts produced by partial melting of the mantle wedge. Enriched, primitive andesites are commonly observed in the western Aleutians because partial melts of subducted sediment and mantle peridotite are much less abundant there. Sediment flux is greatly reduced in the western Aleutians compared to the central arc. Slow convergence rates in the western part of the arc lead to heating of subducting crust and cooling of the mantle wedge. High subduction zone temperatures cause extensive dehydration of sediments and oceanic crust at shallow depths, beneath the forearc region, leaving little H<sub>2</sub>O to flux mantle melting.

Partial melts of subducted eclogite are transported through the coldest, deepest portion of the mantle wedge in cracks or partially molten, pyroxenite diapirs. At shallower depths, they interact with mantle peridotite via reactions which increase melt mass, increase molar Mg#, Ni contents, and MgO contents of derivative liquids, and decrease SiO<sub>2</sub> contents of derivative liquids. The andesitic – rather than basaltic – nature of the derivative liquids in the western Aleutians probably reflects the effect of alkalis and H<sub>2</sub>O, increasing the silica content of olivine-saturated liquids. Where the mantle wedge is hotter and/or more hydrous, similar reactions may increase magma mass still further, producing a less enriched, basaltic liquid product. In this context, partial melts of eclogite could be a ubiquitous “flux” component in many arcs, even where enriched, primitive andesites are not observed.

End-member, enriched primitive andesites in the western Aleutians provide important insights into the nature of eclogite melt components which may be present in other arcs. They

have high Th concentrations, similar to proposed partial melts of sediment in the Marianas, Lesser Antilles and central Aleutian arcs. Thus, high Th, and high Th/Nb, are not good discriminants between eclogite and sediment melt components. However, Th/La correlates well with Pb isotope ratios in Aleutian lavas, so Th/La may be used to distinguish between eclogite and sediment melt components. Interestingly, Th/La also correlates with Ba/La, suggesting that Ba in Aleutian lavas may be transported in partial melts of subducted material rather than, or in addition to, aqueous fluids. Enriched, primitive andesites also have very high Na contents at a given Mg#, compared to other Aleutian lavas. Na concentration correlates with trace element ratios such as Sr/Y and Dy/Yb which are indicative of an eclogite melt component in Aleutian magmas. This suggests that Na is transported in partial melts of subducted eclogite, and cannot be used as an indicator of the degree of melting of the mantle wedge, at least in the western Aleutians.

Moderate amounts of reaction of enriched primitive andesites with upper mantle peridotite, followed by crystal fractionation and mixing with evolved lavas, have produced high Mg# andesite lavas and plutonic rocks with compositions very similar to continental crust. We consider it likely that the processes which form high Mg# andesites in the western Aleutians are similar to those which formed continental crust. In the Archean, the mantle wedge above subduction zones was composed of highly depleted, shallow residues of ~ 40% decompression melting beneath spreading ridges or hotspots. Enriched, primitive andesite magmas were transported through the mantle wedge without extensive dilution with mantle-derived melts because of the highly refractory nature of these residual peridotites.

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research can provide unique and essential constraints on arc processes and continental genesis.

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## Figure Captions

**Figure 1:** Overview of the Aleutian island arc, highlighting the geographic features mentioned in the text, and showing the approximate convergence vector near Adak (N42W) and Attu (N49W) from Engebretson et al. [*Engebretson et al.*, 1985].

**Figure 2:** Histogram illustrating the number of Aleutian lava samples in our compilation as a function of longitude along the arc. Sources of data are given in the caption for Figure 3.

**Figure 3:** Major element variation in Aleutian lavas along the strike of the arc. In this and all subsequent plots of major element oxide variation, weight percent (wt%) concentrations are normalized to 100% volatile free, with all Fe as FeO. Figure 3A illustrates variation in Mg#, or molar Mg/(Mg+Fe), wt% MgO, and calculated magmatic temperature. Magmatic temperatures were calculated using the empirical olivine/liquid thermometer of [*Gaetani and Grove*, 1998], with olivine compositions estimated from liquid compositions using the olivine/liquid Fe/Mg Kd of [*Baker et al.*, 1996], assuming that 80% of Fe is ferrous. Compositions of Aleutian lavas have been previously compiled by Myers [*Myers*, 1988], Kay and Kay [*Kay and Kay*, 1994a], and Kelemen [*Kelemen*, 1995]. More recently, compiled data have been made available by James Myers and Travis McElfrish at <http://www.gg.uwyo.edu/aleutians/index.htm>. The online compilation includes data from the following sources: [*Arculus et al.*, 1977; *Baker and Egger*, 1987; *Brophy*, 1984; *Brophy*, 1986; *Brophy*, 1987b; *Brophy*, 1988; *Brophy*, 1990a; *Brophy*, 1990b; *Brophy*, 1991; *Byers*, 1959; *Coats*, 1952; *Coats*, 1953; *Coats*, 1959; *Coats et al.*, 1961; *Conrad et al.*, 1983; *DeBari et al.*, 1987; *DeLong*, 1974; *DeLong et al.*, 1985; *Drewes et al.*, 1961; *Fenner*, 1926; *Fournelle and Marsh*, 1991; *Fournelle*, 1988; *Franks*, 1981; *Fraser and Barrett*, 1959; *Gust and Perfit*, 1987; *Kay*, 1977; *Kay*, 1978b; *Kay et al.*, 1986; *Kay et al.*, 1978a;

*Kay and Kay, 1985a; Kay and Kay, 1994b; Kay et al., 1982; Marsh, 1976; Marsh, 1981; Marsh, 1982a; Marsh, 1982c; Marsh and Leitz, 1979; McCulloch and Perfit, 1981; McElfresh and Myers, 1994; Merrill, 1885; Miller, 1992; Miller et al., 1992a; Morris, 1984; Morris and Hart, 1983; Myers, 1985; Myers, 1986; Myers, 1987; Myers, 1991; Myers, 1994; Myers and Frost, 1994; Myers et al., ; Nelson, 1959; Neuweld, 1987; Nicaloysen and Myers, 1994; Nye, 1983; Nye and Reid, 1986b; Nye and Reid, 1987; Nye et al., 1986; Perfit, 1977; Perfit, 1983; Perfit et al., 1980a; Romick, 1982; Romick et al., 1990; Simons and Mathewson, 1953; Singer et al., 1992a; Singer et al., 1992b; Snyder, 1959; von Drach et al., 1986; White and Patchett, 1984]. In our current compilation, these have been supplemented with additional data from: [Class et al., 2000; Fraser and Barrett, 1959; Fraser and Snyder, 1956; Gates et al., 1971; Goldstein, 1986; Miller et al., 1994; Rubenstone, 1984; Sun, 1980; Tsvetkov, 1991; Yogodzinski et al., 1995; Yogodzinski et al., 1993; Yogodzinski et al., 1994].*

**Figure 4:** Compositions of Aleutian lavas (A) and plutonic rocks (B) compared to estimates of the composition of continental crust (shaded regions in each plot). Panel (C) shows compositions of lavas and plutonic rocks of the Izu-Bonin-Mariana arc system. Aleutian lavas from Adak and west of Adak, and plutonic rocks throughout the Aleutians, commonly have the SiO<sub>2</sub>, Mg#, K, and La/Yb ratios comparable to estimates for bulk continental crust. Lavas and plutonic rocks of the Izu-Bonin-Mariana system generally do not have these characteristics. Sources of data for Aleutian lavas in caption for Figure 2. Aleutian plutonic rock data from [Citron, 1980; Drewes et al., 1961; Fraser and Barrett, 1959; Fraser and Snyder, 1956; Gates et al., 1971; Kay et al., 1986; Kay et al., 1983; Kay et al., 1990; Perfit et al., 1980b; Romick et al., 1990; Tsvetkov, 1991], and Sue Kay, pers. comm. 2001. Estimates of continental crust composition have been recently compiled and reviewed in [Christensen and Mooney, 1995; Kelemen, 1995; Rudnick, 1995; Rudnick and Fountain, 1995]. Marianas data are from [Bloomer

*et al.*, 1989; *Dixon and Batiza*, 1979; *Elliott et al.*, 1997; *Stern*, 1979; *Stern and Bibee*, 1984; *Wood et al.*, 1981; *Woodhead*, 1989]. Data on the Tanzawa plutonic complex (northern end of the Izu-Bonin-Marianas arc system) are from [*Kawate and Arima*, 1998].

**Figure 5:** Histograms of rock compositions in the Aleutian arc. Top bars show range of estimates for the composition of continental crust. Please note that the histograms for lavas from Adak and west are cumulative, so that the maximum value for each bin refers to the sum of all samples from the Adak area + distal Aleutians + Komandorsky block. Data sources in caption for Figure 3. Section 1.2 and Figure 1 give definitions of regional terms.

**Figure 6:** MORB-normalized spidergrams for estimated compositions of the continental crust (Figure 3A) and Aleutian lavas. In the Aleutian lava plots, the field of compiled estimates for the continental crust is shown in gray for reference. In Figure 3A, arrows emphasize elements in continental crust which may be concentrated in detrital sediments (Nb, Ta in rutile, Zr, Hf in zircon) or removed in solution during surficial weathering (Sr). In general, Aleutian magmas have higher La/Nb, La/Ta, Sm/Zr, and Sm/Hf and lower Nd/Sr than continental crust. Other than that, we feel that the similarities in trace element abundance between Aleutian lavas and continental crust is striking, and indicates that similar processes operated during formation of continental crust and Aleutian magmas. Data sources in caption for Figure 3. MORB normalization values from [*Hofmann*, 1988]. Note that Plank and Langmuir [*Plank and Langmuir*, 1998] suggested that some estimates of Nb and Ta concentration in the continental crust [*McLennan and Taylor*, 1985; *Rudnick and Fountain*, 1995] are too high by a factor of ~ 2. However, because these estimates provide neither an upper nor a lower bound in Figure 3A, we have not modified the published values.

**Figure 7:**  $^{206}\text{Pb}/^{204}\text{Pb}$ ,  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $^{143}\text{Nd}/^{144}\text{Nd}$  in Aleutian lavas versus longitude, illustrating systematic along-strike variation in these ratios. Data sources in caption for Figure 3, plus new Pb isotope data for western Aleutian lavas in Table 1.

**Figure 8:** Some major element abundances in Aleutian lavas versus trench orthogonal convergence rate beneath Aleutian volcanoes. Convergence rate from [Fournelle *et al.*, 1994] is plotted versus the longitude of each volcano. However, the convergence rate for each volcano is determined at a position in the trench that is updip of the volcano, projected along the plate convergence vectors shown in Figure 1 [Engebretson *et al.*, 1985]. Longitudes of positions in trench updip of each volcano were calculated using the following empirical function:

$$\text{longitude (}^\circ\text{W)} = -393.71 + 5.6627 * \text{volcano longitude (}^\circ\text{W)} - 0.013863 * \text{volcano longitude (}^\circ\text{W)}^2$$

**Figure 9:** Geologic section along the Aleutian trench based on [McCarthy and Scholl, 1985; Ryan and Scholl, 1989; Scholl *et al.*, 1987; Vallier *et al.*, 1994; von Huene, 1986] and our unpublished data, sediment thickness subducted beneath Aleutian volcanoes, based on this section, sediment flux beneath Aleutian volcanoes (product of convergence rate from Figure 8 and sediment thickness), and Pb isotope ratios versus sediment flux. Sources of Pb isotope data in caption for Figure 3, and new data in Table 1.

**Figure 10:** Variation of major element oxides versus molar Mg# in Aleutian lavas. Sources of data in caption for Figure 3.

**Figure 11:** Isotope variation in Aleutian lavas and in sediments from DSDP Site 183 (52.57°N, 161.20°W) southeast of the Aleutian trench. Sources of data in caption for Figure 3, [Peucker-

*Ehrenbrink et al.*, 1994; *Plank and Langmuir*, 1998; *von Drach et al.*, 1986], new Pb isotope data for western Aleutian lavas in Table 1, and new Sr and Nd isotope data for DSDP Site 183 sediments in Table 2.

**Figure 12:** Proposed trace element indicators of a recycled sediment component in arc lavas, versus longitude along the Aleutian arc. Of these, only Th/La and perhaps Ba/La seem to show systematic variation along-strike. Sources of data in caption for Figure 3.

**Figure 13:** Variation of  $^{207}\text{Pb}/^{204}\text{Pb}$ , Th/La and Ba/La in Aleutian lavas, and Th/La versus Ba/La for a compilation of arcs worldwide – excepting the Aleutians – from Elliott (this volume). In the Aleutians, Th/La appears to be correlated with  $^{207}\text{Pb}/^{204}\text{Pb}$ . Th/La is definitely correlated with Ba/La, as previously noted by [*Kay*, 1980; *Kay and Kay*, 1994a; *Yogodzinski et al.*, 1994]. In the “worldwide” data set, Ba/La is negatively correlated with Th/La. High Ba/La has been interpreted to reflect addition of an aqueous fluid component to the arc magma source, whereas high Th/La has been interpreted to reflect addition of a distinct, partial melt of subducted sediment. These components cannot be easily separated in the Aleutians. Sources of Aleutian data in caption for Figure 3.

**Figure 14:** Proposed trace element indicators of an aqueous fluid component in arc lavas, versus longitude along the Aleutian arc. Sources of data in caption for Figure 3.

**Figure 15:**  $^{207}\text{Pb}/^{204}\text{Pb}$  versus Ce/Pb in Aleutian lavas. Closed circles indicate data from volcanoes at and west of Adak. All closed circles that are not labeled with a location are from Adak Island. This plot is similar to Figure 4 of [*Miller et al.*, 1994], but includes data for volcanoes other than Okmok and Rechesnoi, plus updated fields for DSDP Site 183 sediments

and MORB. The plot demonstrates that relatively high Ce/Pb is common in Aleutian lavas with unradiogenic Pb isotopes. Sources of Aleutian data in caption for Figure 3, plus new Pb concentrations and Pb isotope ratios for western Aleutian lavas in Table 1. Field for MORB glasses from the East Pacific Rise and Juan de Fuca Ridge based on data in the RIDGE petrological database at <http://petdb.ldeo.columbia.edu/>. Field for DSDP Site 183 sediments from data in [Miller *et al.*, 1994; Peucker-Ehrenbrink *et al.*, 1994; Plank and Langmuir, 1998; von Drach *et al.*, 1986]. Field for “aqueous fluid” is based on the reasoning of [Brenan *et al.*, 1995a; Miller *et al.*, 1994], but expanded to include fluid derived from subducted sediment as well as altered MORB.

**Figure 16:** Sr/Y, a proposed indicator of an eclogite melt component, versus other compositional data for Aleutian lavas. Sources of data in caption for Figure 3. Where Y data were not available, Sr/(9.6\*Yb) was used instead, based on the chondritic Y/Yb ratio. For REE, (N) indicates that concentrations are normalized to C1 chondrite values [Anders and Grevesse, 1989].

**Figure 17:** Sr/Y and chondrite normalized REE ratios in Aleutian lavas versus longitude. Sources of data in caption for Figure 3. C1 chondrite normalization factors from [Anders and Grevesse, 1989]. Where Y data were not available, Sr/(9.6\*Yb) was used instead, based on the chondritic Y/Yb ratio.

**Figure 18:** Variation in Dy/Yb(N) (and Tb/Yb), Th/La, Ce/Pb and Ba/La in Aleutian lavas. Sources of data in caption for Figure 3, plus new Pb concentrations for western Aleutian lavas in Table 1.

**Figure 19:** Calculated trends for mixing of (a) primitive basaltic lavas from Okmok volcano [Nye and Reid, 1986a] with (b) rhyolite ia32 (also called 47ABY28) from Rechesnoi volcano [Byers, 1959; Kay and Kay, 1994a].

**Figure 20:** Dy/Yb(N) (and Tb/Yb) versus molar Mg# for Aleutian lavas, illustrating that the most primitive lavas – with the highest Mg# – have the highest Dy/Yb (and Sr/Y and La/Yb, Figure 16). This is inconsistent with the mixing trajectories in Figure 19, involving primitive lavas with flat REE patterns mixing with evolved lavas with enriched REE patterns. Instead, mixing of primitive, enriched lavas with evolved lavas having flat REE patterns is consistent with the Aleutian data. Sources of data in caption for Figure 3.

**Figure 21:** Histogram of wt% SiO<sub>2</sub> for primitive Aleutian lavas. Sources of data in caption for Figure 3.

**Figure 22:** Concentration of Th versus Pb in primitive Aleutian lavas. Sources of data in caption for Figure 3.

**Figure 23:** Results of trace element modeling. Grey field shows estimated bulk composition of continental crust, from Figure 6A with data sources in caption. Crystal/liquid distribution coefficients, model rock compositions, and data sources for these values are given in Table 4.

Figure 23A illustrates compositions of aggregated liquids formed by fractional melting [Gast, 1968; Shaw, 1970] of garnet peridotite (pyrolite). Figure 23B illustrates compositions of aggregated liquids formed by fractional melting of eclogite (MORB). Figure 23C illustrates liquid compositions resulting from reaction of MORB with spinel peridotite (pyrolite) to produce garnet pyroxenite, using the AFC equations of [DePaolo, 1981], with the mass of solid

reactants/mass of solid products set to 0.97. The ratio of initial liquid mass/integrated mass of peridotite reactant ranges from ~0.3 at a melt fraction of 0.9 to ~0.03 at a melt fraction of 0.1.

Partial melting of eclogite produces a trace element pattern similar to some enriched, primitive andesites in the Aleutians, but with much lower heavy REE than in continental crust. Partial melting of garnet peridotite, and reaction producing garnet pyroxenite, produce steep REE patterns in resulting liquids, but without the fractionations of Nb/La, Ce/Pb and Sr/Nd seen in Aleutian lavas and continental crust.

Figure 23D shows liquids resulting from reaction of a small degree melt of eclogite with spinel peridotite (pyrolite), with no change in solid phase proportions and the mass of solid reactants/mass of solid products,  $M_a/M_c$ , of 1.05. The melt/rock ratio (initial mass of liquid/integrated mass of peridotite reactant) in these models ranges from ~0.2 where the liquid mass is 1.25 times its initial mass to ~0.02 where the liquid is 3x initial mass. Figure 23E shows results of reaction between the same eclogite melt and depleted spinel peridotite (abyssal peridotite), with no change in solid phase proportions,  $M_a/M_c$  of 1.02, and melt/rock ratio from ~0.1 to 0.01. Figure 23F shows results of reaction between the same eclogite melt and depleted peridotite, with no change in solid phase proportions,  $M_a/M_c$  of 1.04, and melt/rock ratio from ~0.02 to 0.002, extending to very large increases in liquid mass. Crystal/liquid distribution coefficients and model rock and liquid compositions are given in Table 4.

Moderate amounts of reaction between partial melts of eclogite and spinel peridotite, under conditions of increasing liquid mass, produce liquids with trace element patterns similar to continental crust. Only Zr and Hf in continental crust are not well fit by these models. Large extents of reaction between eclogite partial melts and spinel peridotite produce liquids with nearly flat REE patterns that retain enrichments in Ba, Th, U, and K and fractionations of Nb/La, Ce/Pb and Sr/Nd seen in arc lavas worldwide.

These models are presented to show *possible* outcomes of various igneous processes. We make no claim that these are unique explanations for the trace element contents of Aleutian lavas or continental crust. Also, the models could be incorrect if we have chosen inappropriate crystal/liquid distribution coefficients (Table 4). For the most part, small variations in distribution coefficients resulting from mineral and melt composition in basaltic systems are insignificant for our purposes in modeling. However, there is a substantial difference between distribution coefficients between minerals and basaltic melt, on the one hand, and minerals and andesitic to dacitic melt on the other hand. Whereas mineral/basalt partitioning has been extensively investigated, mineral/dacite partitioning is less well studied. In compiling mineral/dacite distribution coefficients for modeling partial melting of eclogite, we have relied mainly on studies of trace element partitioning between phenocrysts and matrix in volcanic rocks. For some elements (Th, U, Nb, Pb in clinopyroxene/dacite melt) we have had to interpolate values based on the pattern of distribution coefficients for mineral/basalt, and better known values for mineral/dacite for other elements (Ba, K, Ce, Nd in clinopyroxene/dacite melt). For highly incompatible elements, such as Ba, Th, U and K, our interpolated values are not crucial to the results of modeling. However, in considering our model results which show fractionation of Ce/Pb and Sr/Nd, the choice of distribution coefficients could be crucial. Thus, readers should be cautious in evaluating these results. With this said, we note that our models are consistent with the experimental data of [Rapp *et al.*, 1999], and the trace element models of Tatsumi ([Tatsumi, 2000; Tatsumi, 2001], and pers. comm., 2000).

Also, Ti is not a trace element in processes involving residual rutile. As a result, strictly speaking, it should not be modeled here. However, we adjusted the distribution coefficients for Ti in clinopyroxene, garnet and rutile to provide results that are consistent with the experimental data of [Rapp *et al.*, 1999], which show no fractionation of Ti from REE such as Dy and Tb.

**Figure 24:** Difference between observed wt% TiO<sub>2</sub> in primitive Aleutian lavas (molar Mg# > 0.6) and TiO<sub>2</sub> concentration required for rutile saturation. The rutile saturation surface estimated by [Ryerson and Watson, 1987] varies with magma composition, temperature and pressure. Pressures are given in each panel. Magmatic temperatures for each lava composition, with both 0 and 10 wt% H<sub>2</sub>O, were calculated using the olivine/liquid thermometer of [Gaetani and Grove, 1998], with olivine compositions estimated from liquid compositions using the olivine/liquid Fe/Mg Kd of [Baker et al., 1996], assuming that 80% of Fe is ferrous.

**Figure 25:** Schematic illustration of the mantle flow pattern and temperature distribution beneath an arc. In the mantle flow figures, the thickness of the flow vectors is proportional to the flow velocity. In the top two panels, flow in the mantle wedge is driven mainly by coupling with the subducting plate. Such a flow pattern is likely for mantle wedge viscosities greater than ~ 10<sup>19</sup> Pa s. In the bottom two panels, flow in the wedge is driven mainly by thermal convection. Such a pattern is likely for mantle wedge viscosities less than ~ 10<sup>19</sup> Pa s.

**Figure 26:** Schematic phase diagrams for natural silicate melts, projected from spinel onto the normative plane olivine - clinopyroxene - SiO<sub>2</sub> at 2 to 3 GPa (e.g., [Kelemen et al., 1992; Kinzler, 1997]). In both anhydrous, alkali-poor and hydrous, alkali-rich systems, eclogite melts may solidify – forming pyroxenite and undergoing “thermal death” – if they react with a large mass of peridotite at constant temperature.

**Figure 27:** Schematic illustration of the effect of combined heating and decompression on ascending melt in the mantle. Both heating and decompression take the melt further from its liquidus surface, giving it the potential to dissolve wall rock, increasing magma mass. As shown in Figure 26, if a partial melt of eclogite reacts with the mantle at constant temperature, it may

solidify (light gray arrow). If a partial melt of eclogite reacts with the mantle slowly, while heating and decompressing rapidly, it may move along its solid saturation surface (dark gray arrow), or even remain undersaturated in solid phases (black arrow) for part of its ascent path [Kelemen, 1986; Kelemen, 1990; Kelemen, 1995; Kelemen *et al.*, 1993]. A similar illustration, for peridotite melts rising higher in the mantle wedge, is given in [Grove *et al.*, 2001].

**Figure 28:** Possible effects of H<sub>2</sub>O and alkali enrichment during reaction of eclogite melt and mantle peridotite at constant temperature. Where mica and amphibole are not stable, for example for temperatures greater than ~ 1100°C, reactions that decrease magma mass will increase H<sub>2</sub>O and alkali concentration in the remaining melt. These changes will tend to lower the magmatic liquidus surface, and enhance the stability of olivine relative to pyroxene in the local melt-rock system. Ultimately, this could produce small amounts of H<sub>2</sub>O- and alkali-rich, olivine saturated melt, avoiding “thermal death”.

**Figure 29:** Possible modes of melting and melt transport for generation of enriched, primitive andesites and – by analogy – of continental crust. A. Transport by flow in melt-filled fractures. B. Transport in diapirs of hybrid melt + pyroxenite, modified from [Ringwood, 1974]. In both A and B, small amounts of H<sub>2</sub>O-saturated melt form within subducting oceanic crust in eclogite facies, and rise into overlying mantle peridotite. Ascent rates are slow, and reaction creates abundant amphibole pyroxenite and consumes much of the melt. This forms a low permeability barrier (e.g., [Kelemen and Aharonov, 1998]) beneath which additional melt accumulates.

In A, at a high melt fraction, the buoyancy force created by accumulated melt beneath the permeability barrier becomes sufficient to fracture the overlying rock. The fracture propagates until it encounters a magma chamber or melt-rich zone in the uppermost mantle beneath the arc.

(This melt-rich zone in the shallow mantle may also result from an overlying permeability barrier (e.g., [Kelemen and Aharonov, 1998])). Here, melt reacts more extensively with mantle peridotite, and also may mix with partial melts of peridotite.

In B, at a high melt fraction, the mixture of pyroxenite and magma overlying the subducting plate becomes sufficiently buoyant to rise diapirically through the mantle. Heating and decompression increase the melt fraction within the rising diapir, rendering it increasingly permeable, so that melt from within reacts with surrounding peridotite.

In both A and B, orthopyroxene-rich peridotite reaction products mixed with depleted peridotite ultimately accrete to form cratonic upper mantle [Kelemen *et al.*, 1998; Kelemen *et al.*, 1993; Kesson and Ringwood, 1989; Rudnick *et al.*, 1994]. Hybrid melts, with trace element characteristics inherited from eclogite melting and major elements greatly modified by reaction with the mantle, ultimately pass through the Moho as enriched, primitive andesite.

Crystallization of pyroxenite cumulates in the uppermost mantle and at the base of the crust [DeBari and Coleman, 1989; Miller and Christensen, 1994; Müntener *et al.*, 2001], and/or mixing with evolved lavas produced by basalt fractionation [Yogodzinski and Kelemen, 1998], produces high Mg# andesite magmas which rise to form mid-crustal batholiths. Subsequent delamination of pyroxenite cumulates [Arndt and Goldstein, 1989; Jull and Kelemen, 2001; Kay and Kay, 1991; Kay and Kay, 1993; Tatsumi, 2000] leaves a plutonic crustal section with the major and trace element composition of high Mg# andesite and continental crust.

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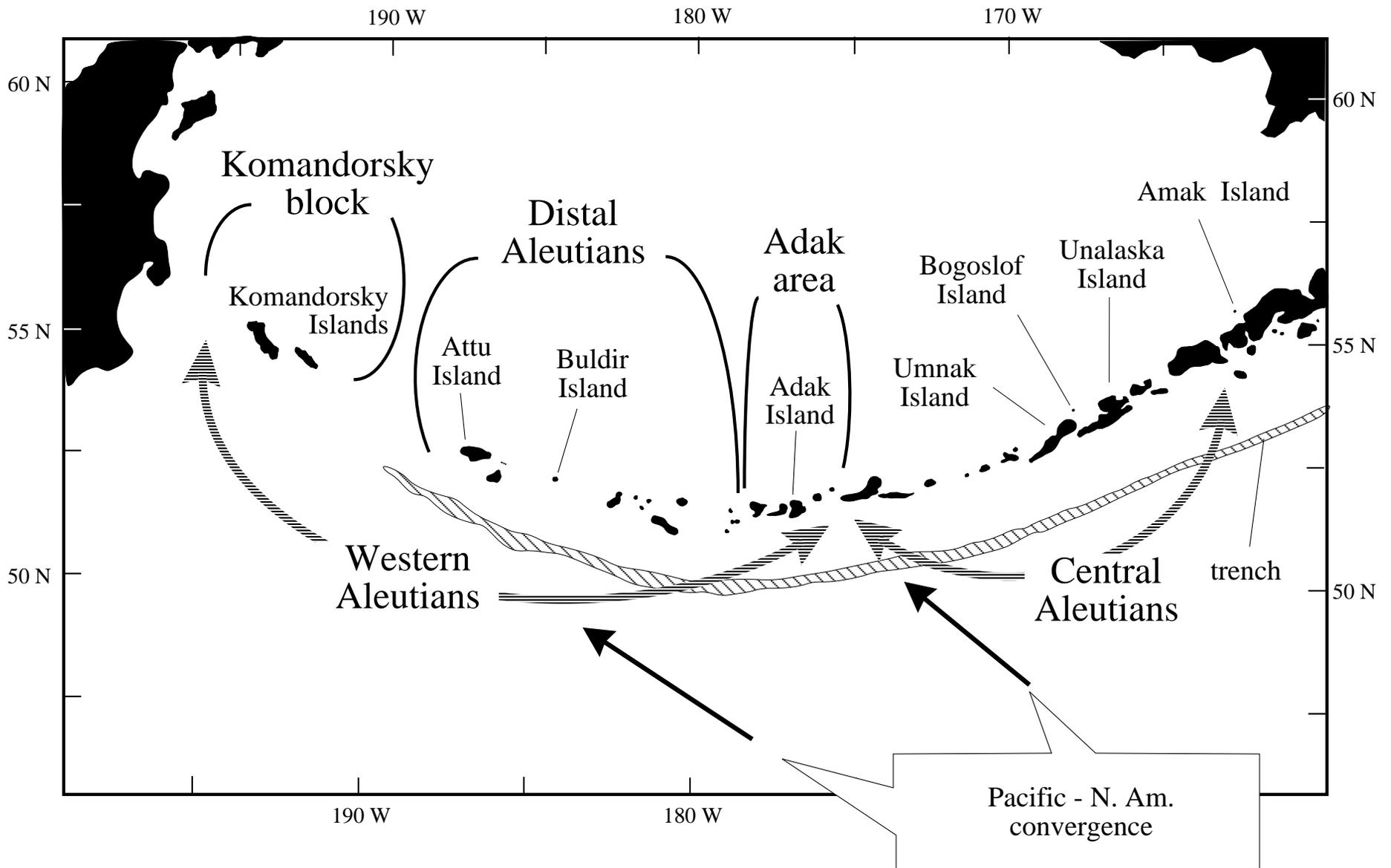
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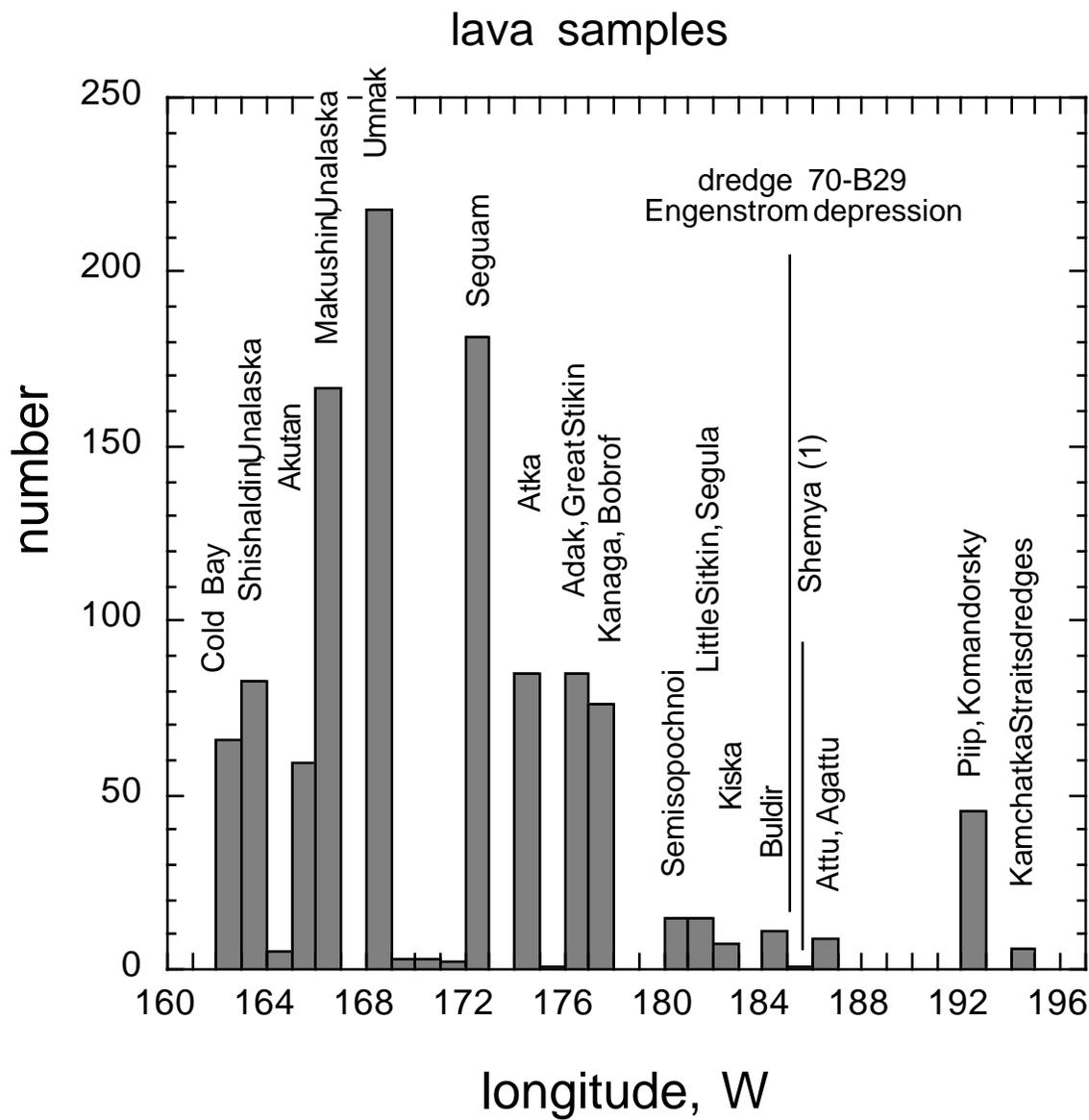
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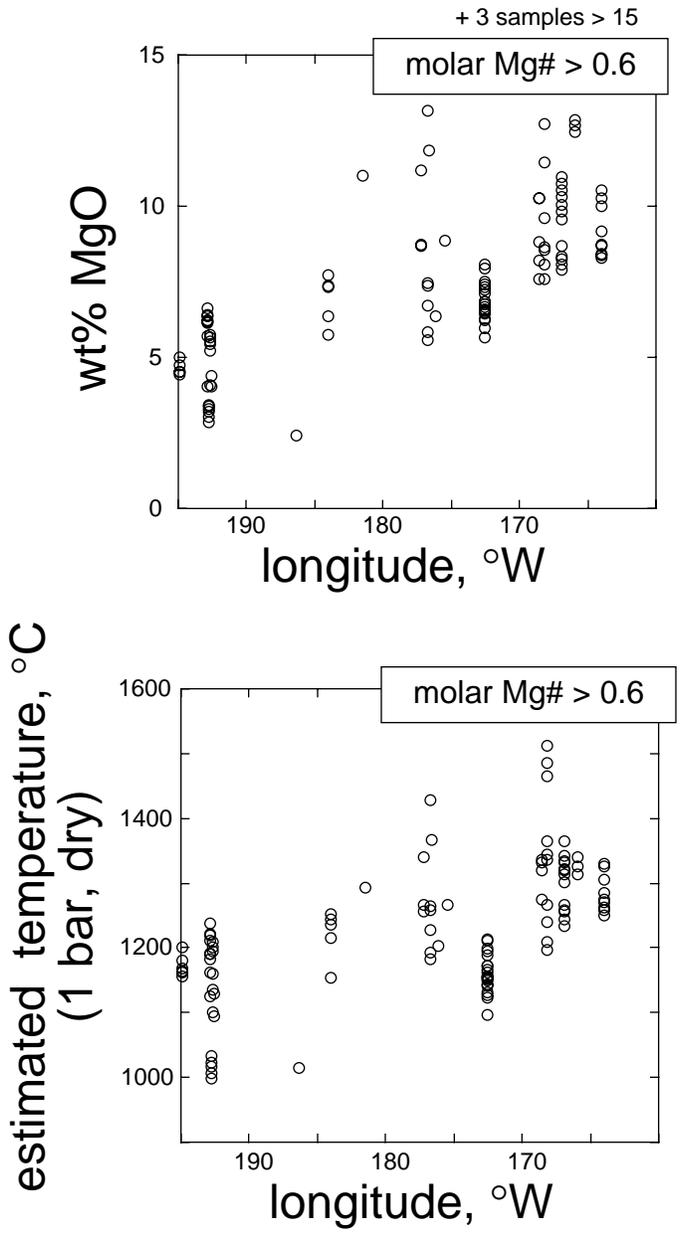
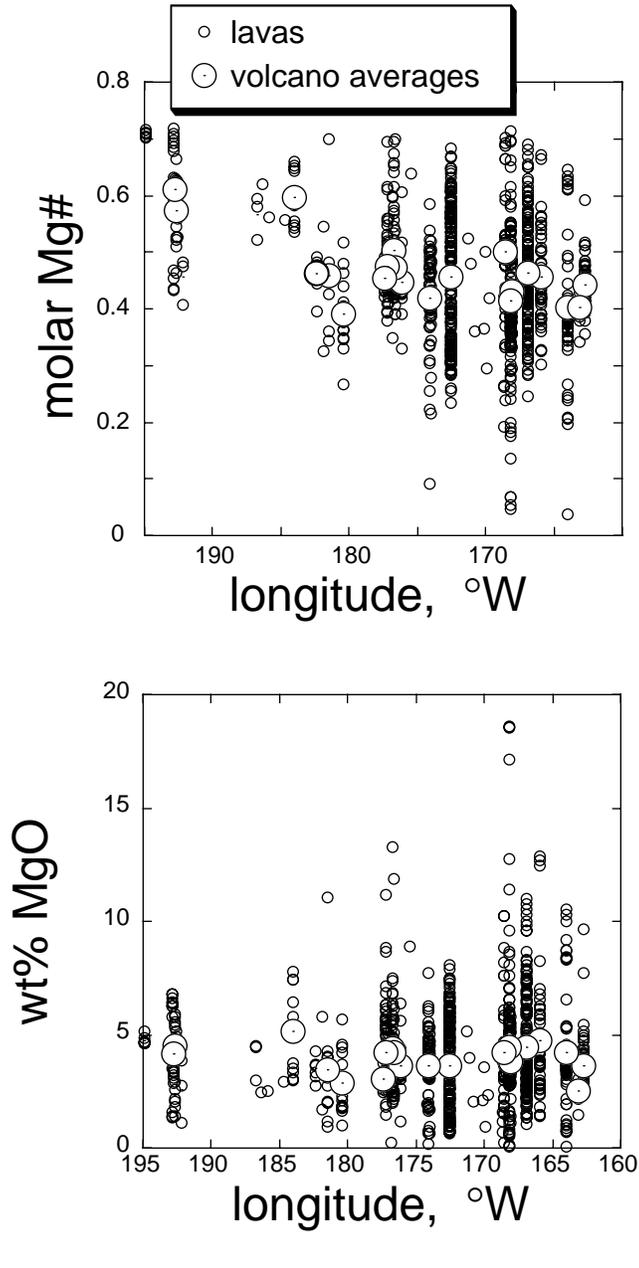
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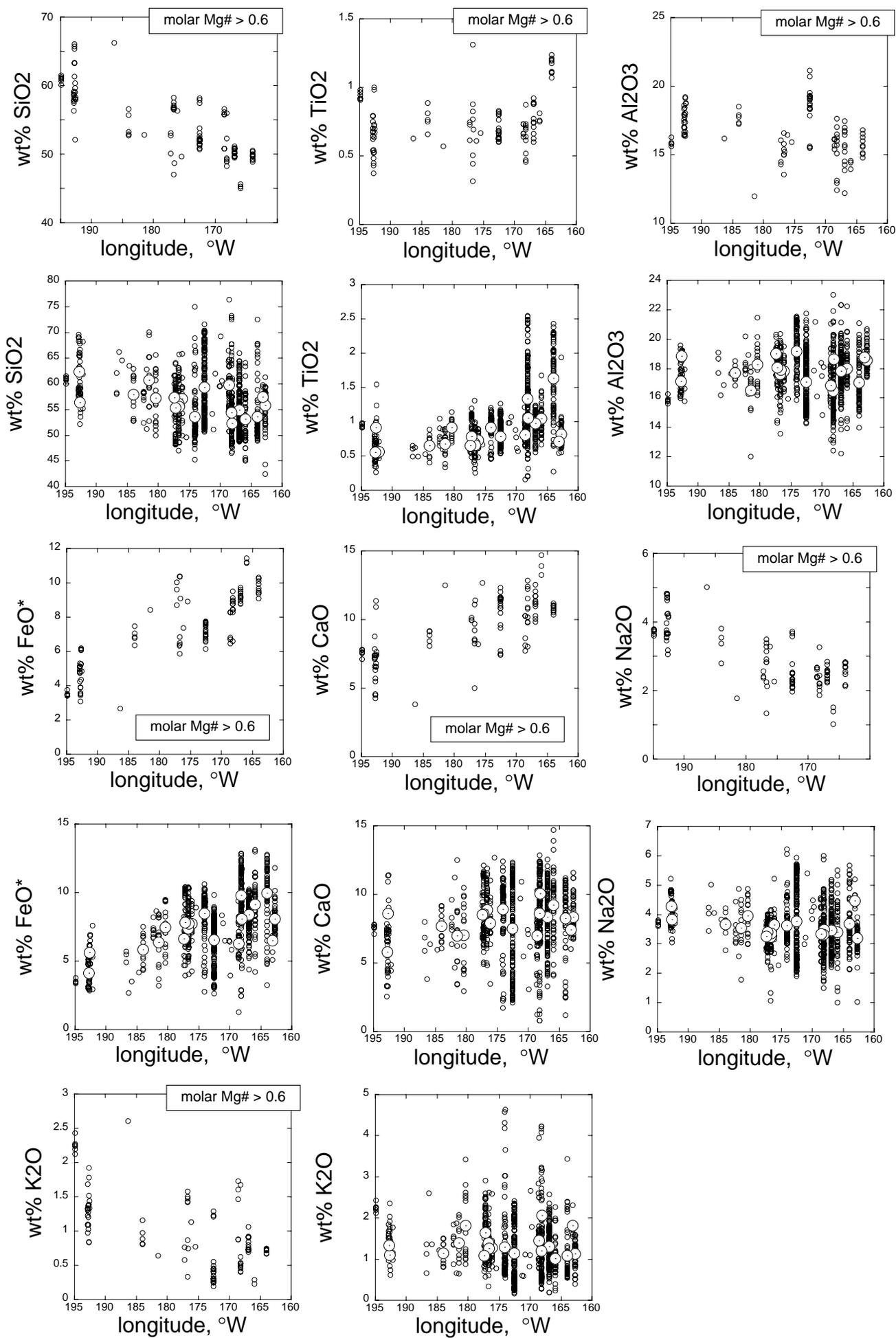


Kelemen et al., 2001, figure 1

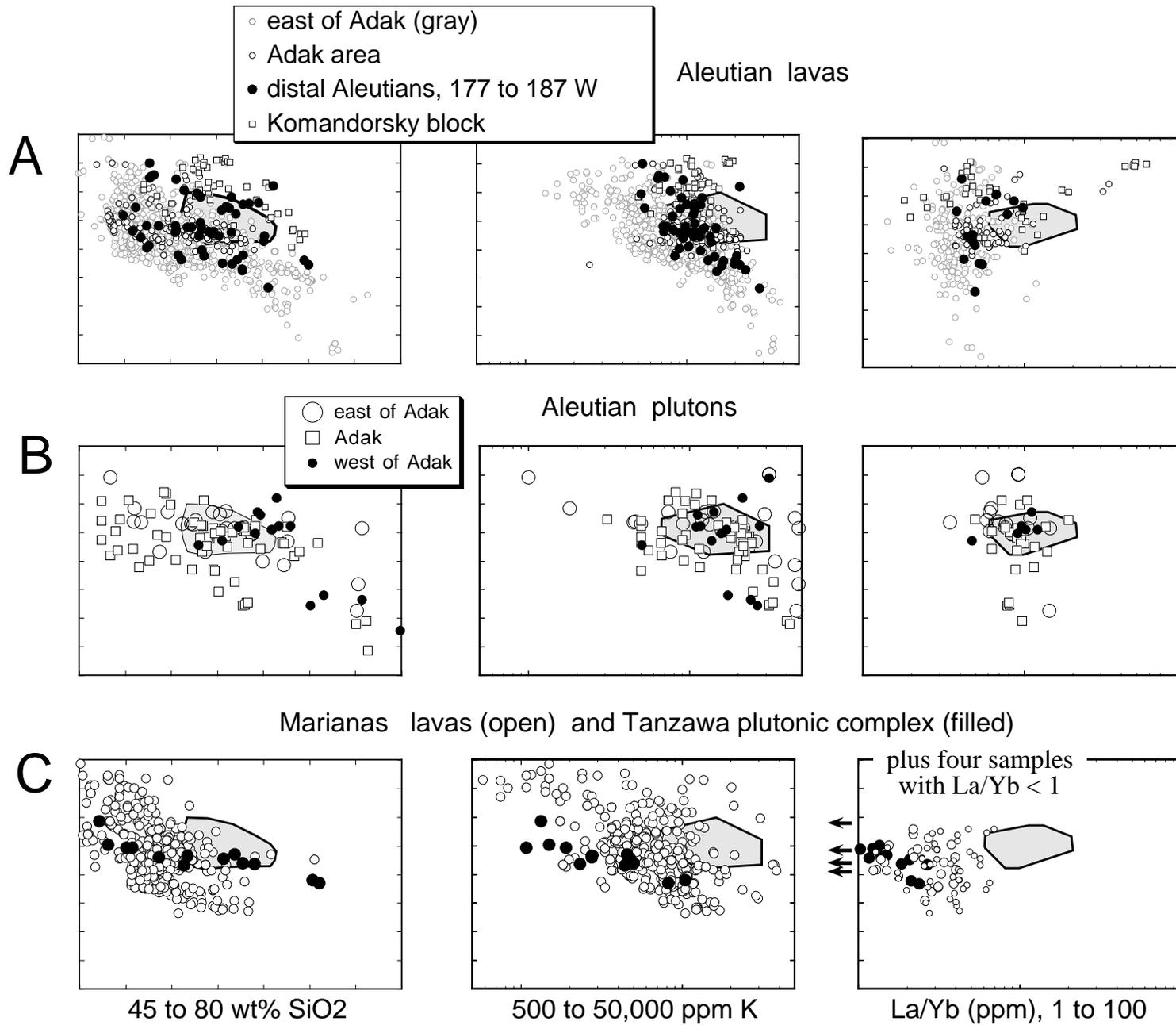


Kelemen et al., 2001, figure 2

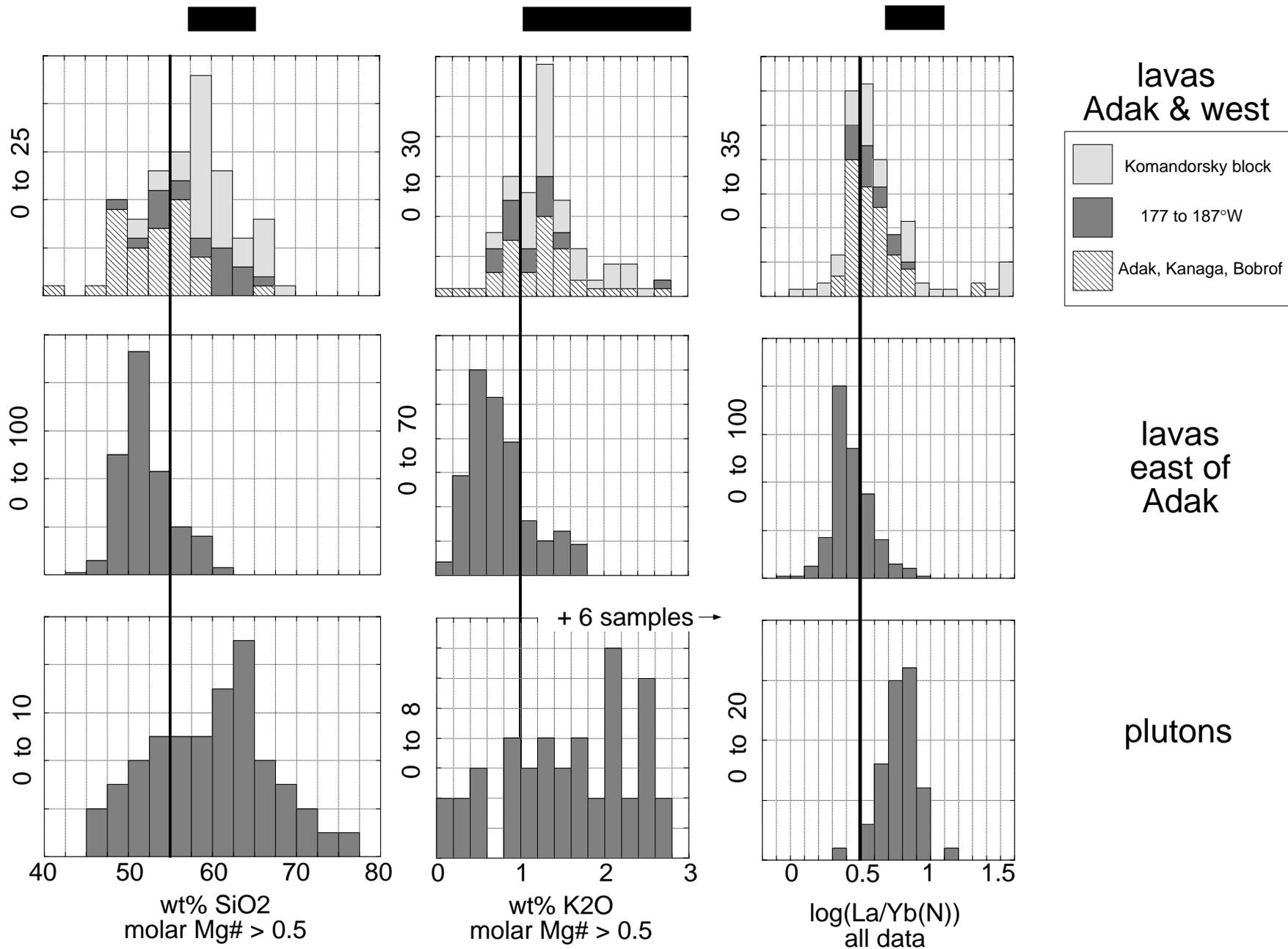




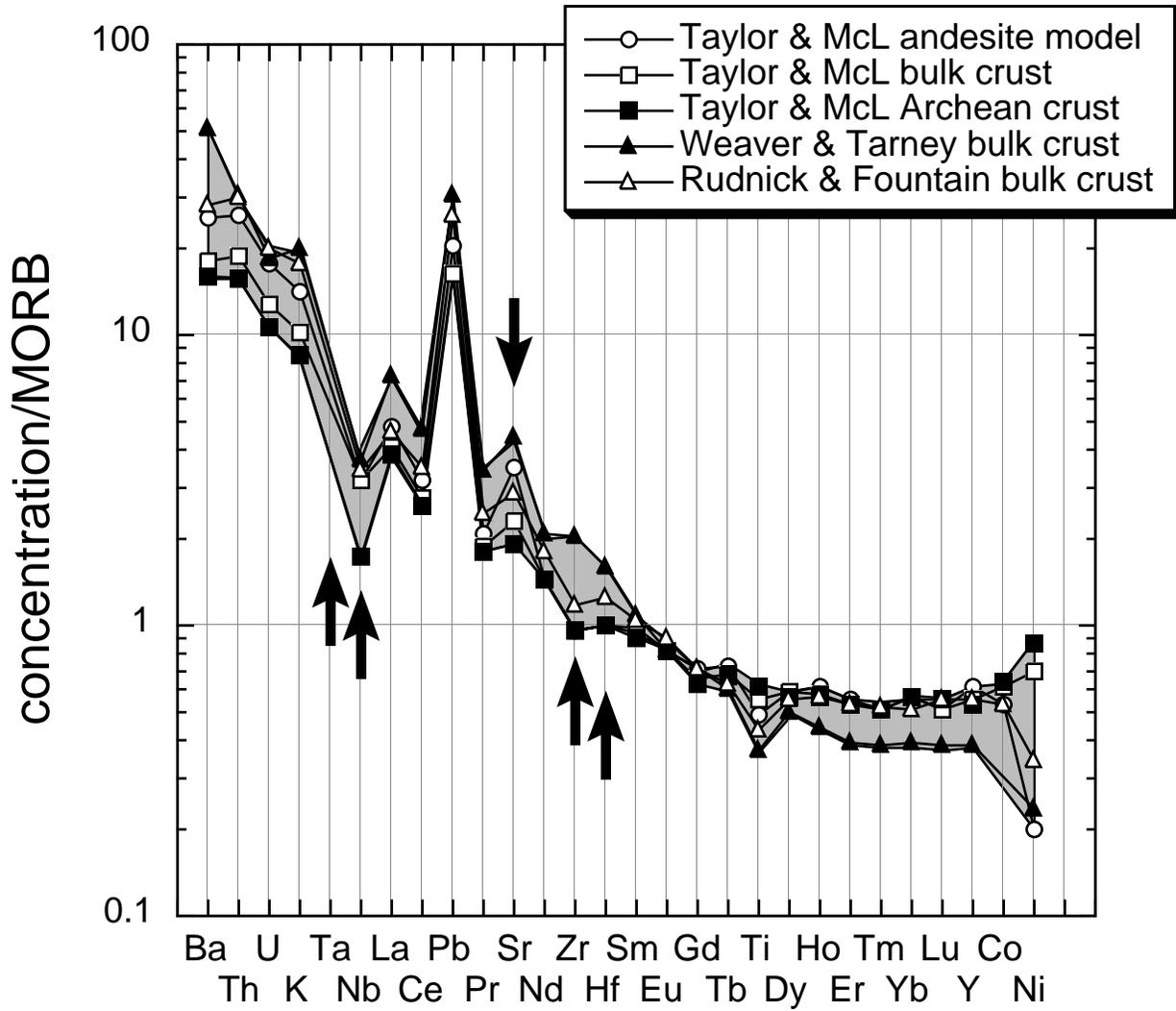
all plots: molar Mg#, 0 to 0.8



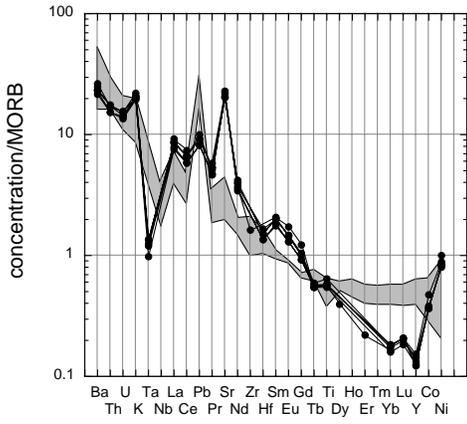
← estimated continental crust (Mg# 0.45 to 0.54) →



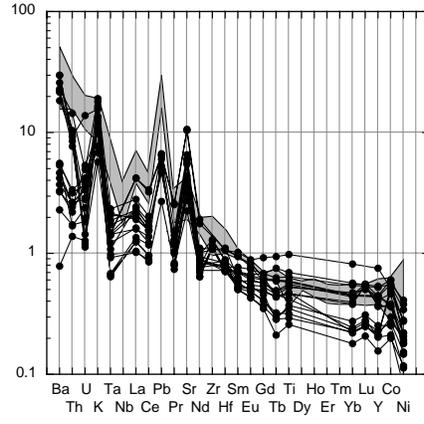
# continental crust



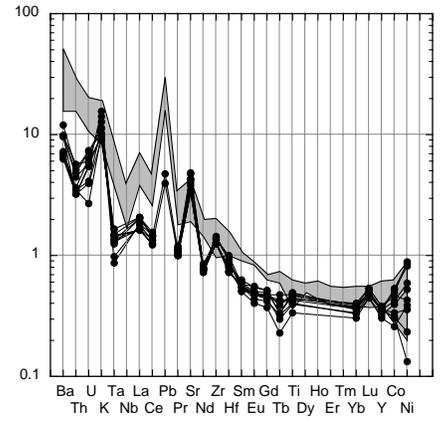
Komandorsky Straits



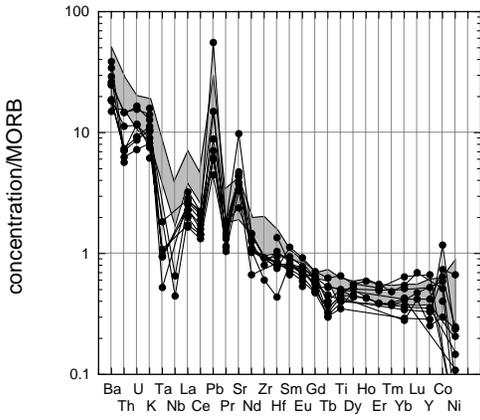
Komandorsky Islands



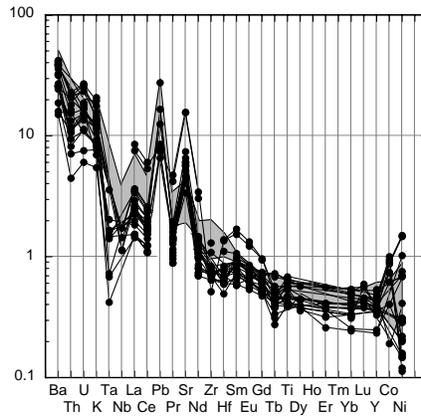
Piip volcano  
Komandorsky block



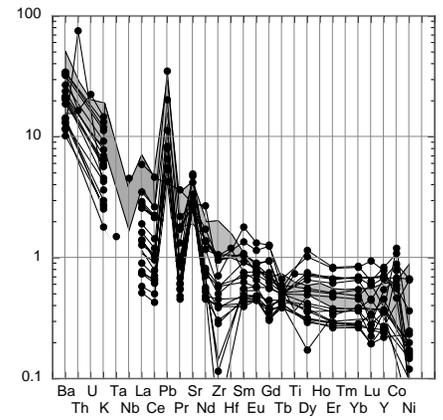
distal Aleutians



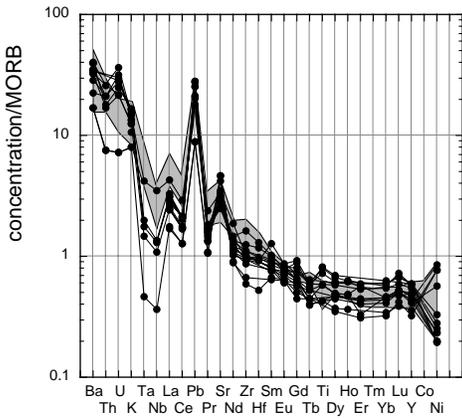
Adak area



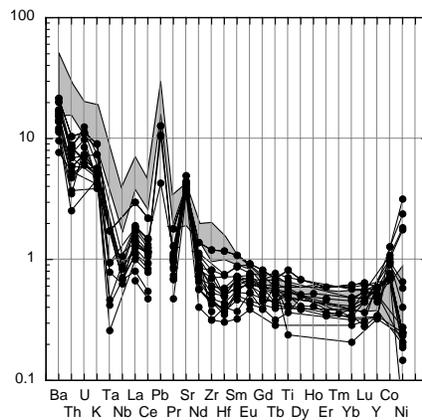
Atka & Seguam



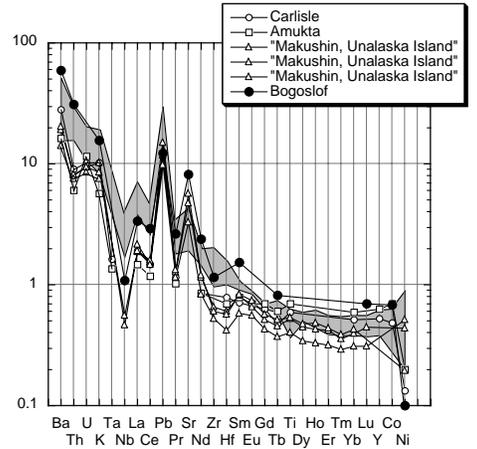
Rechesnoi volcano  
Umnak Island

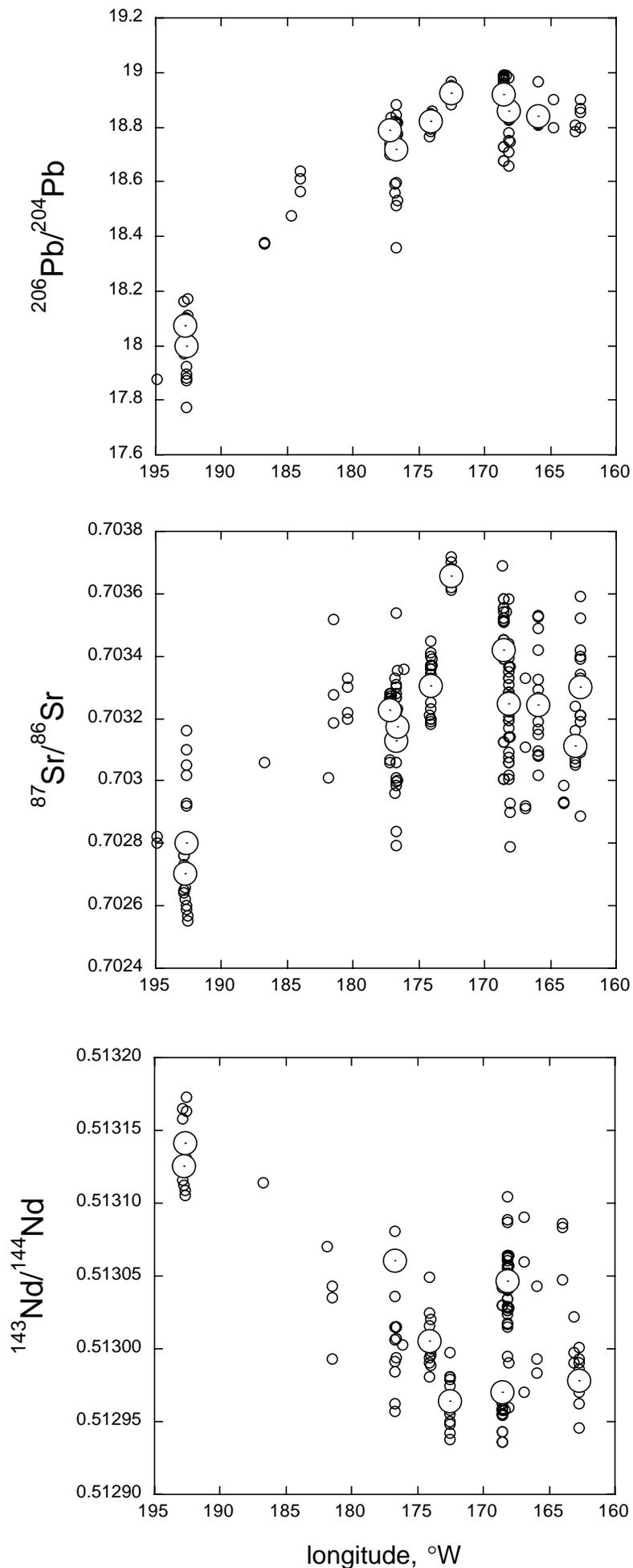


Okmok volcano  
Umnak Island

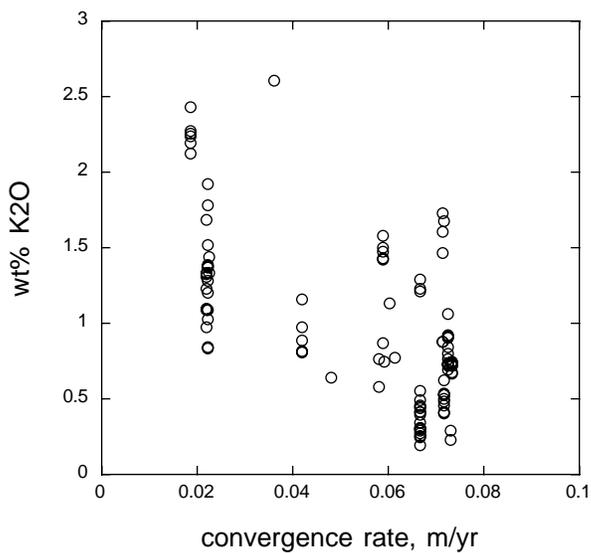
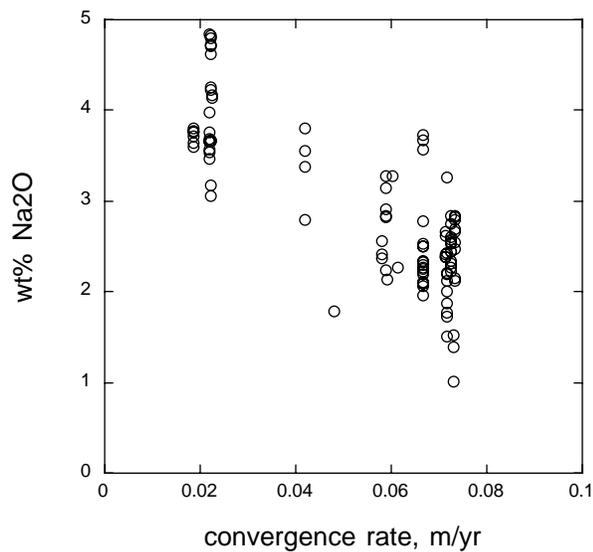
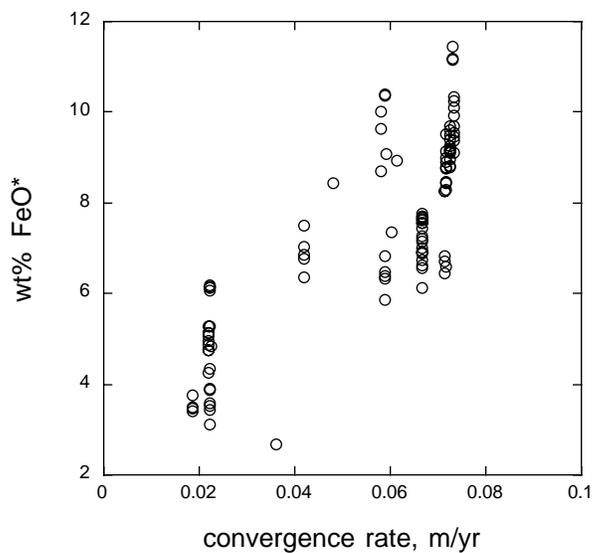
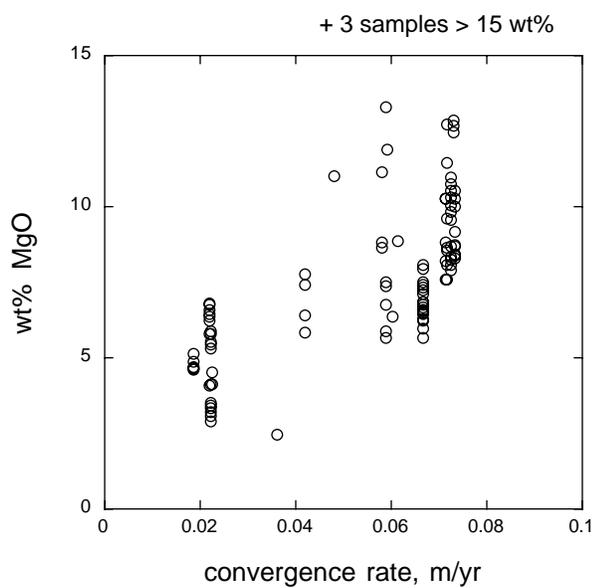
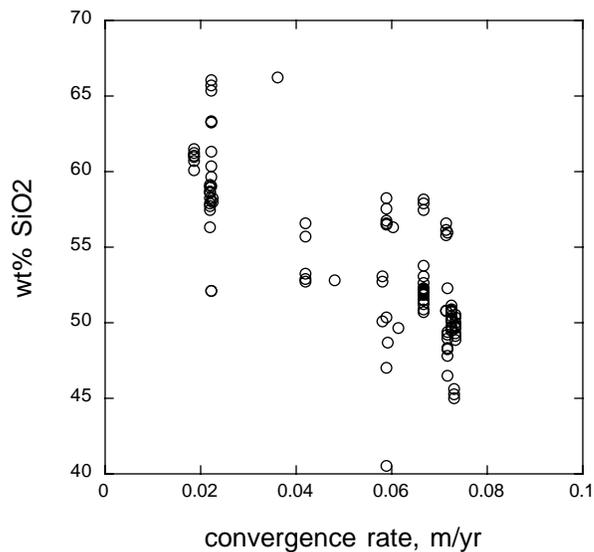
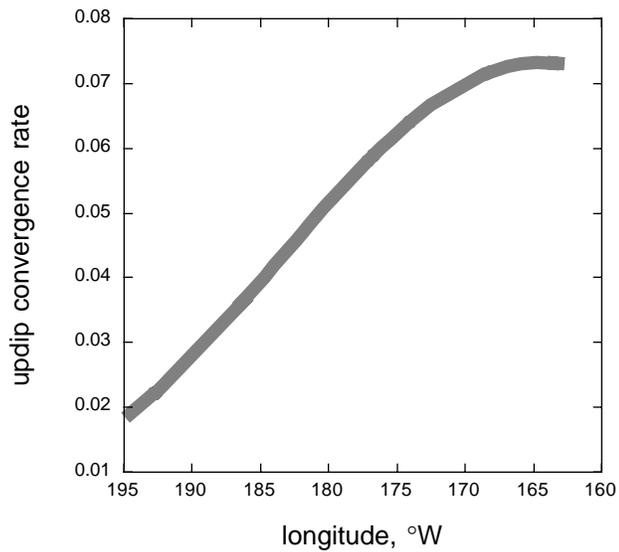


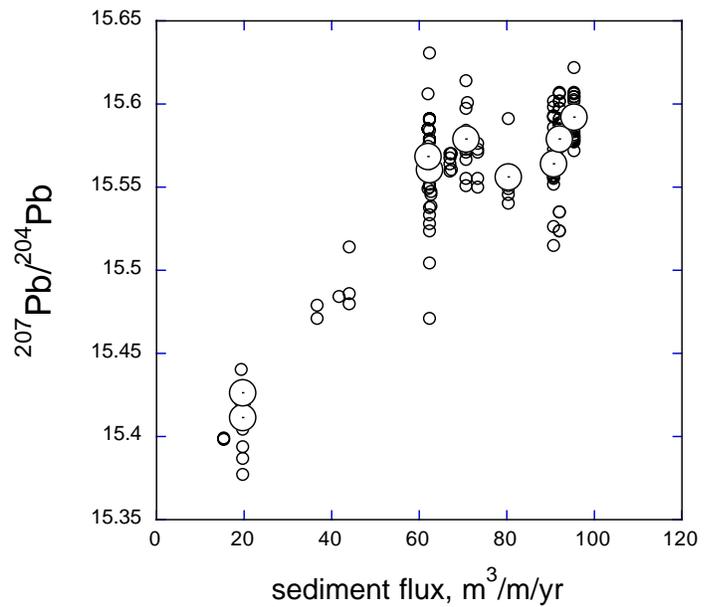
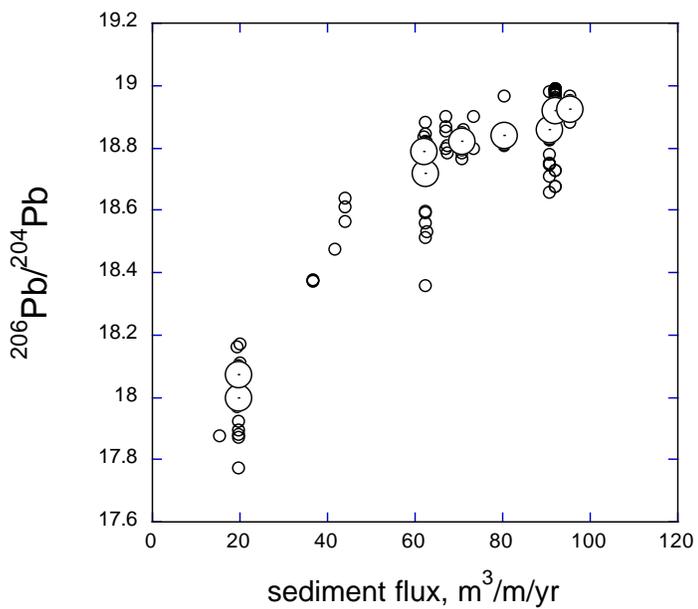
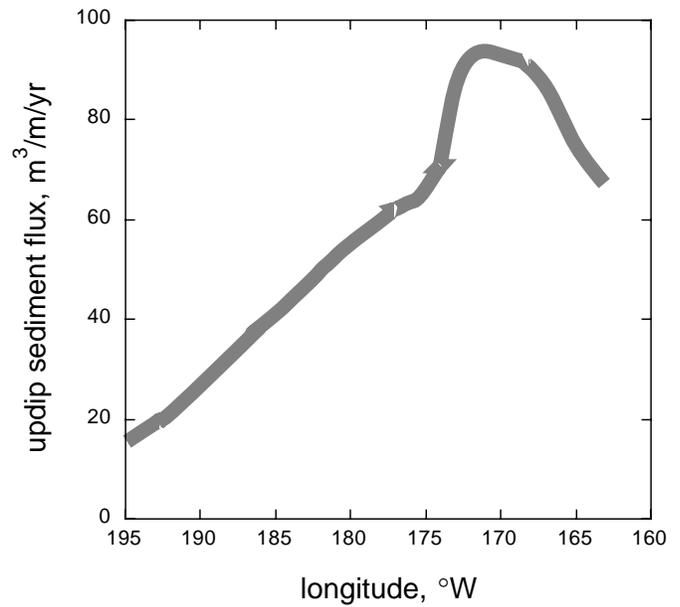
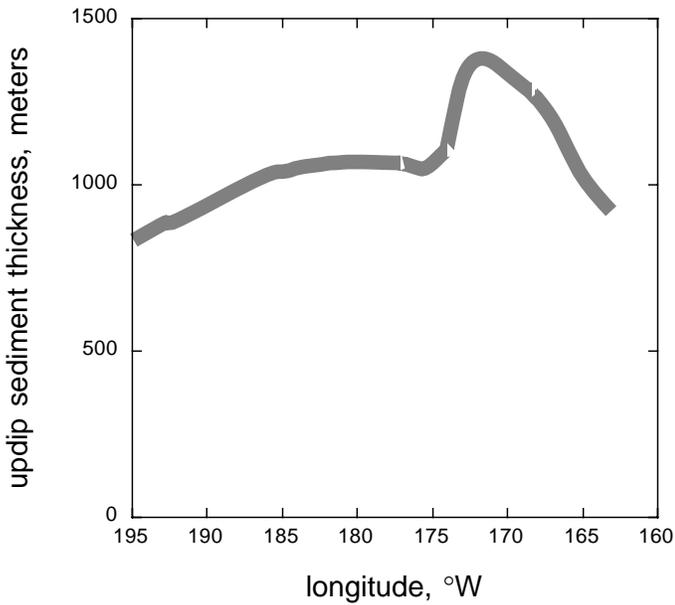
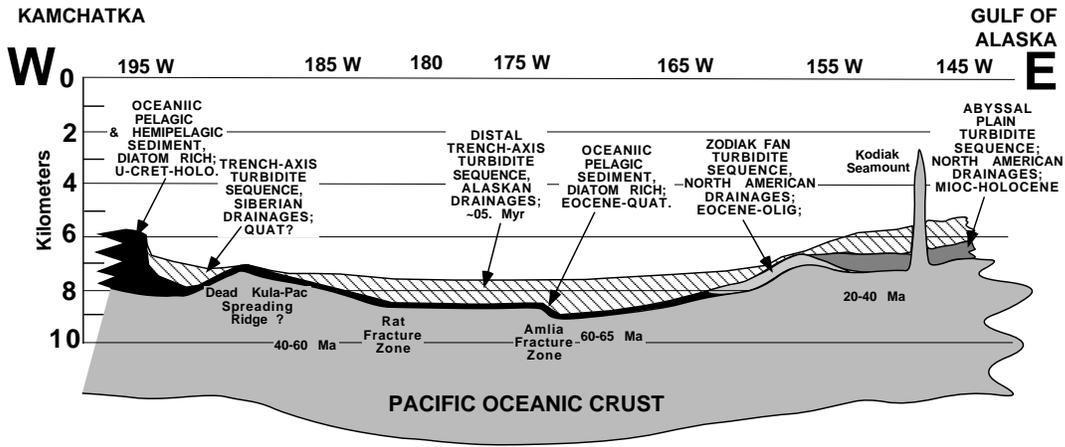
east central Aleutians

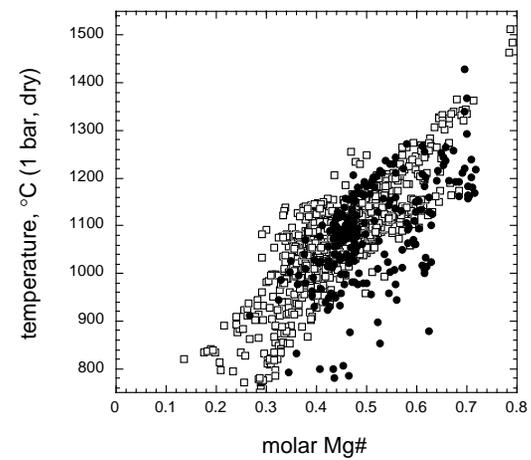
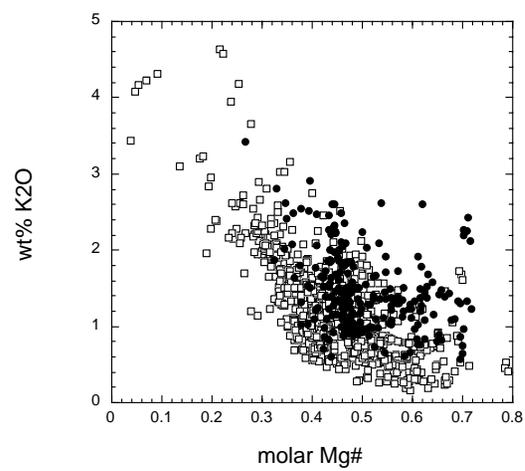
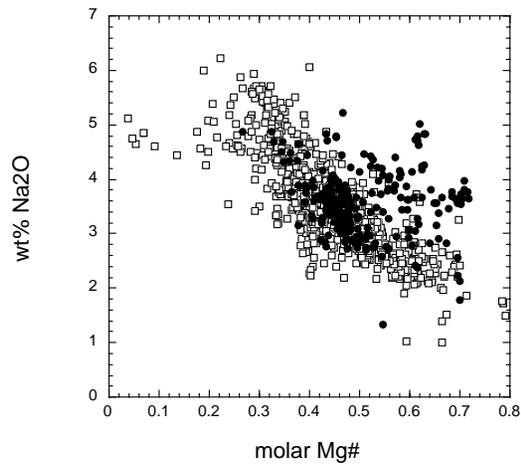
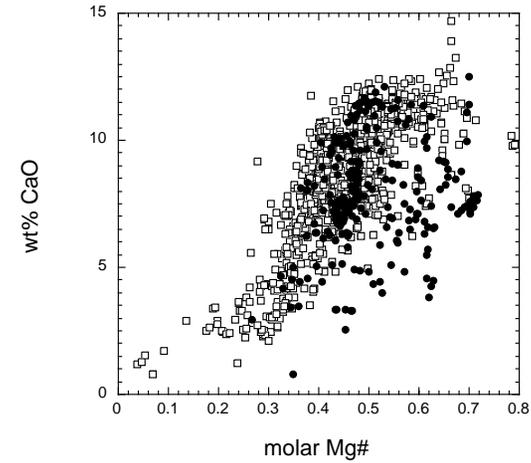
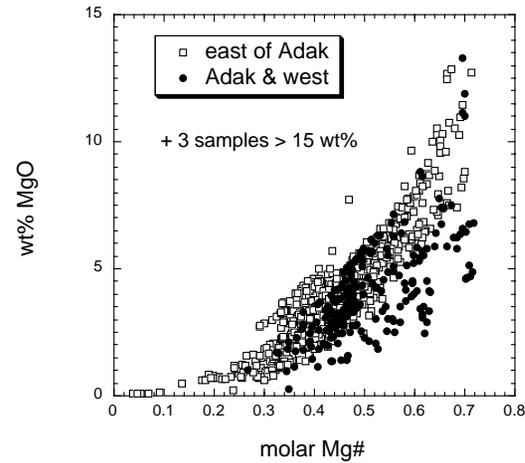
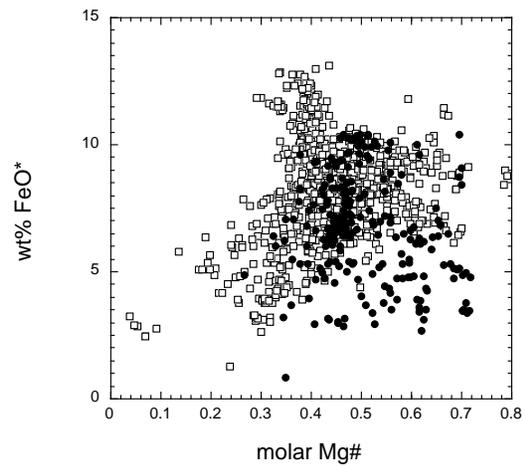
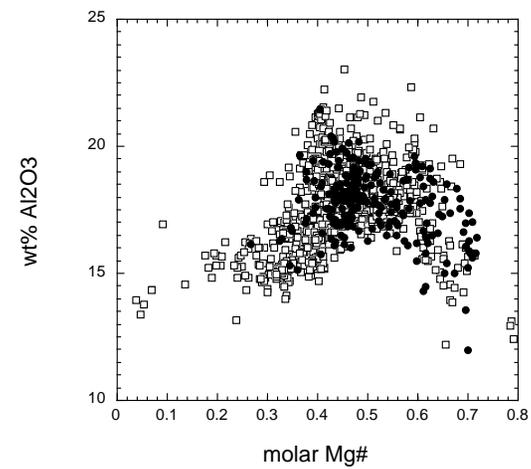
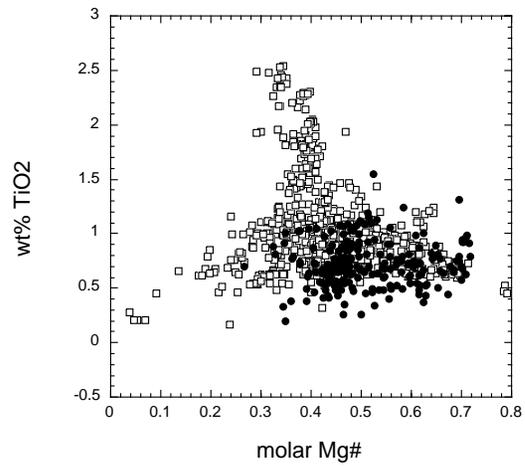
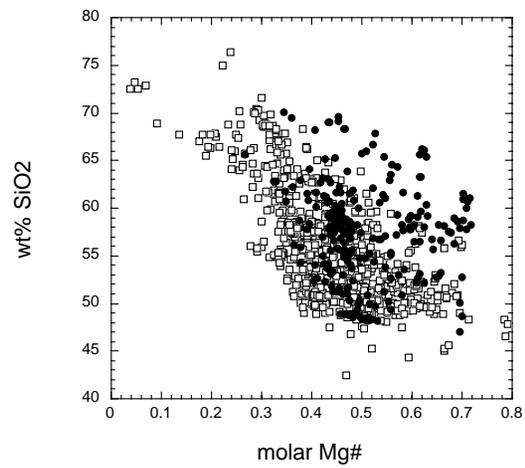


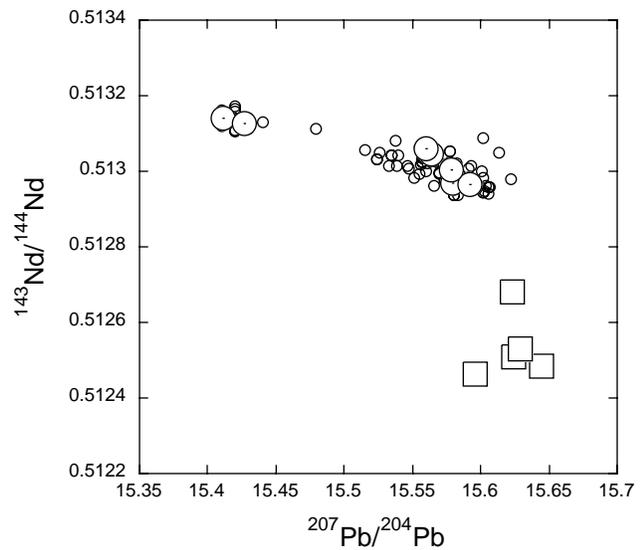
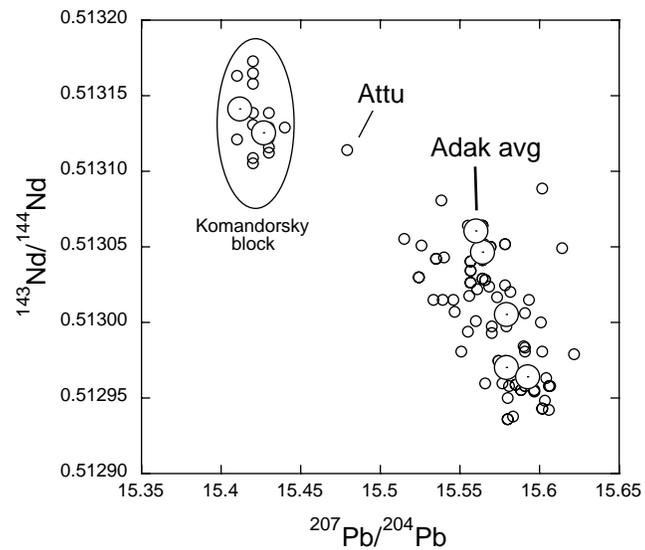
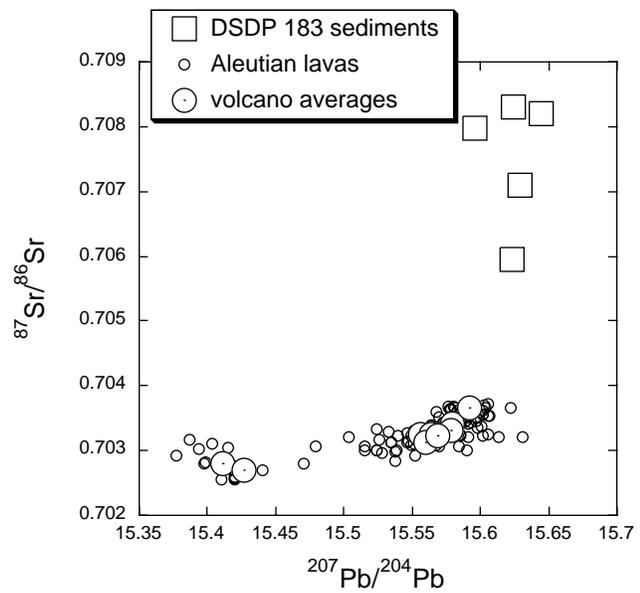
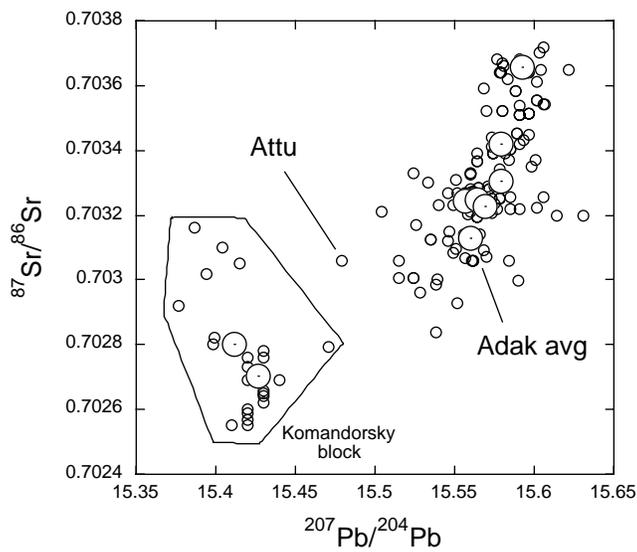
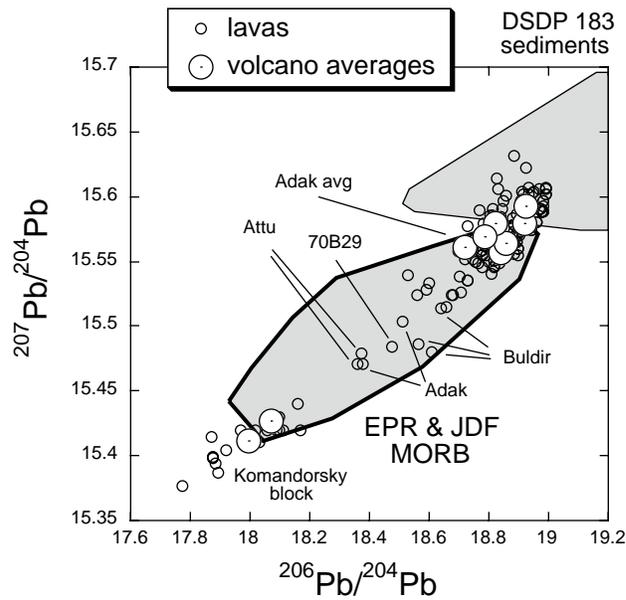


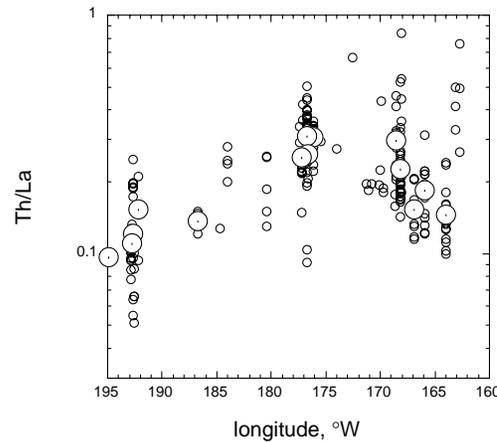
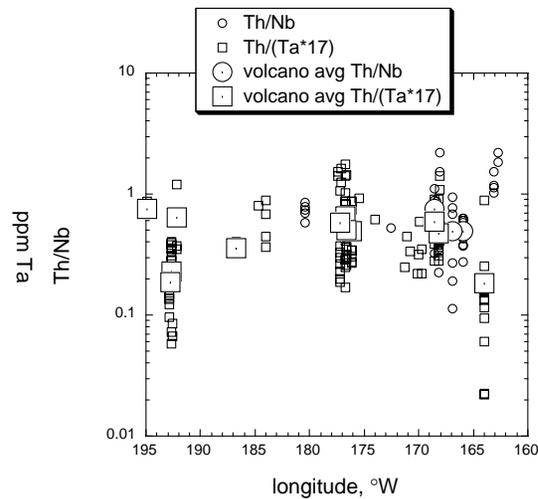
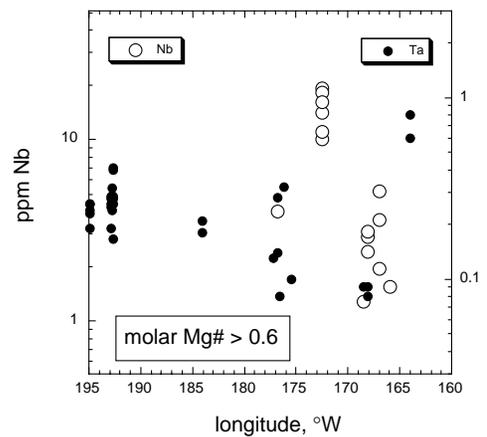
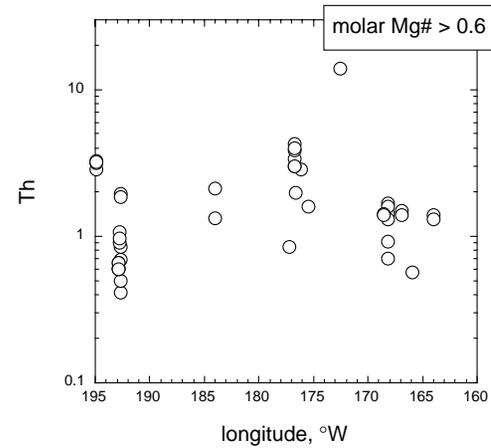
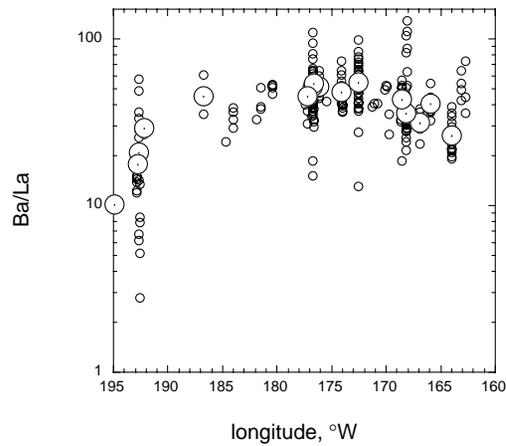
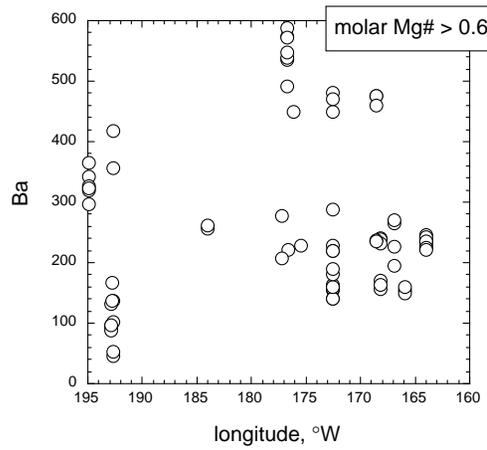
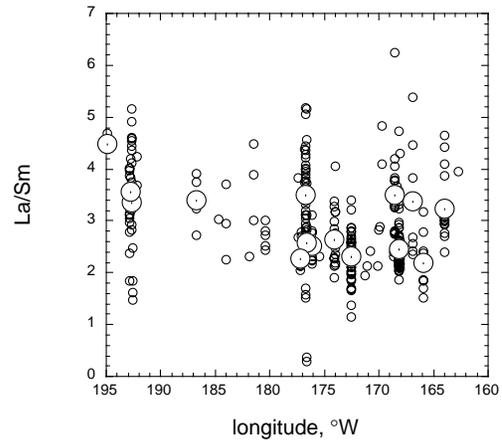
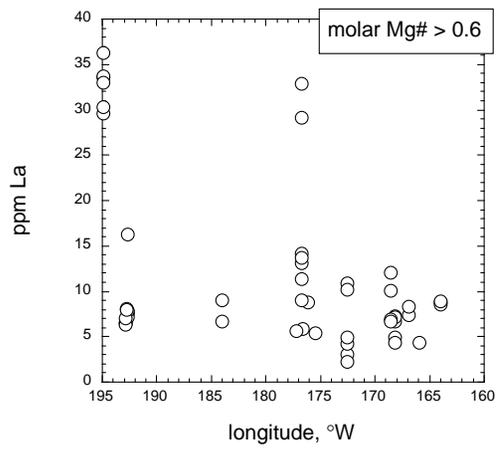
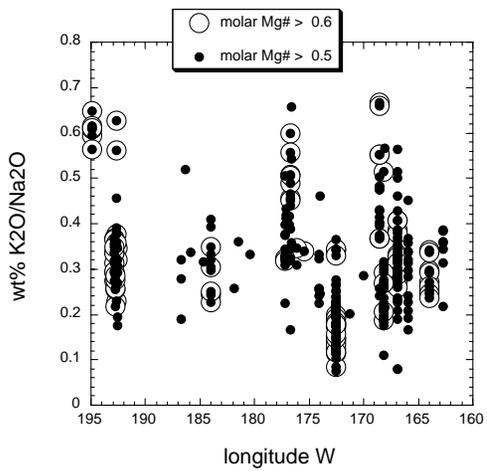
Kelemen et al., 2001, figure 7

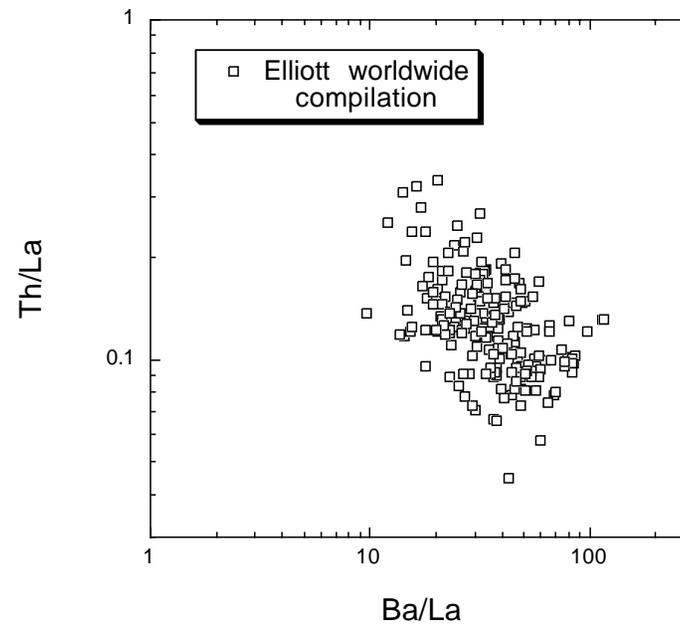
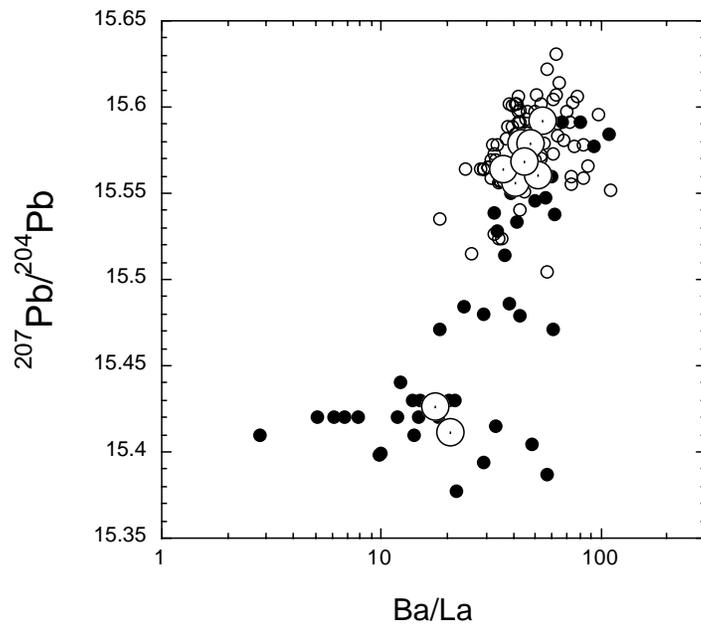
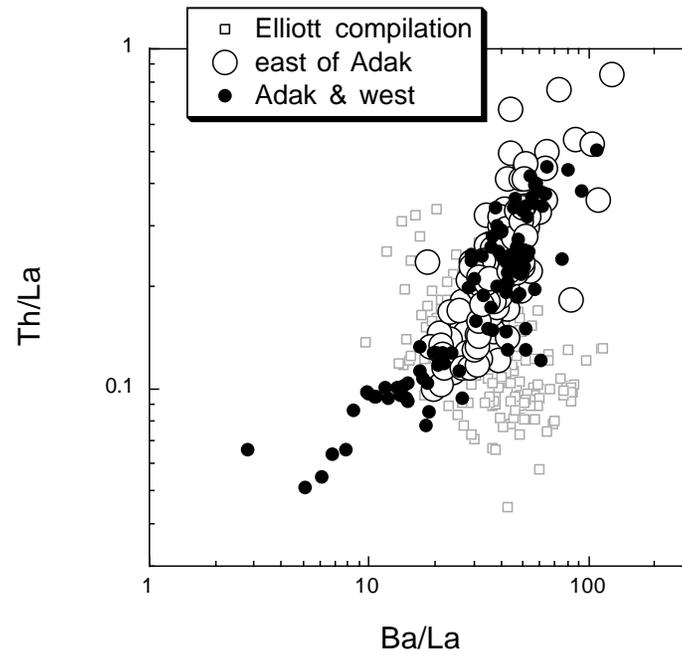
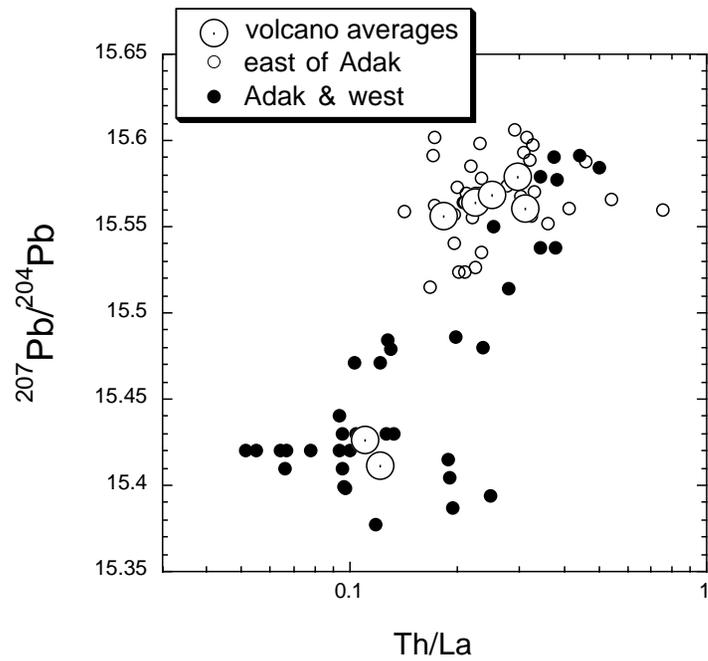


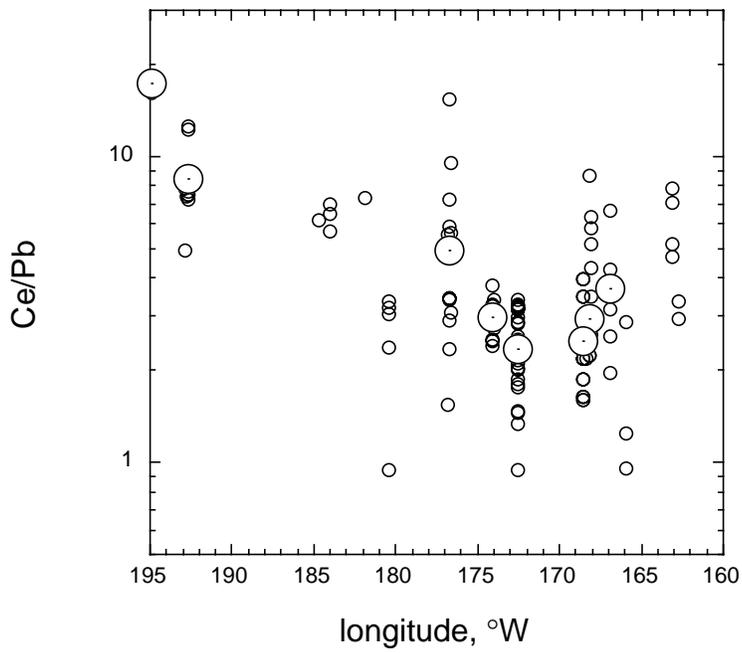
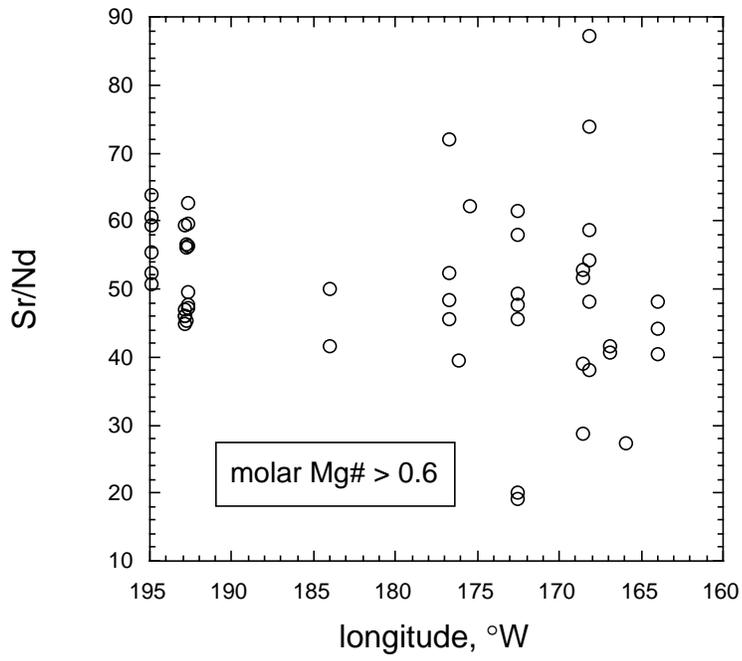
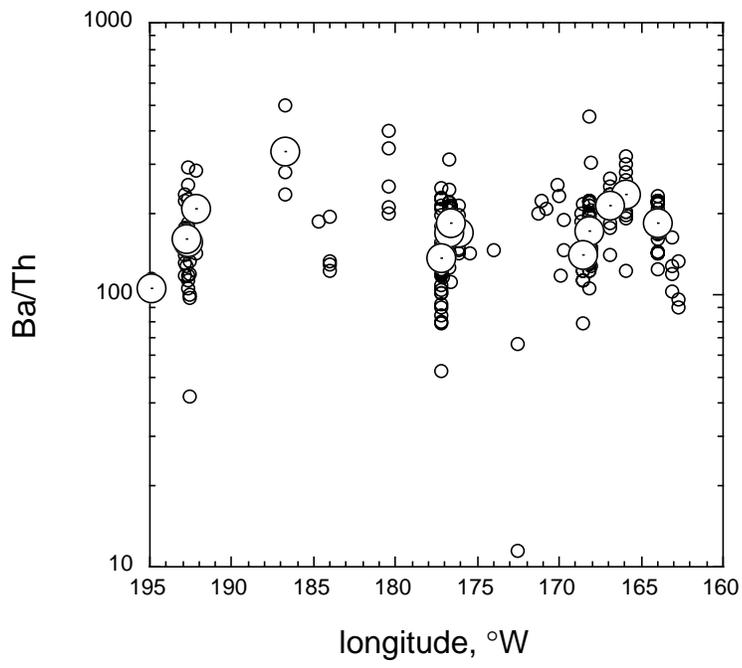


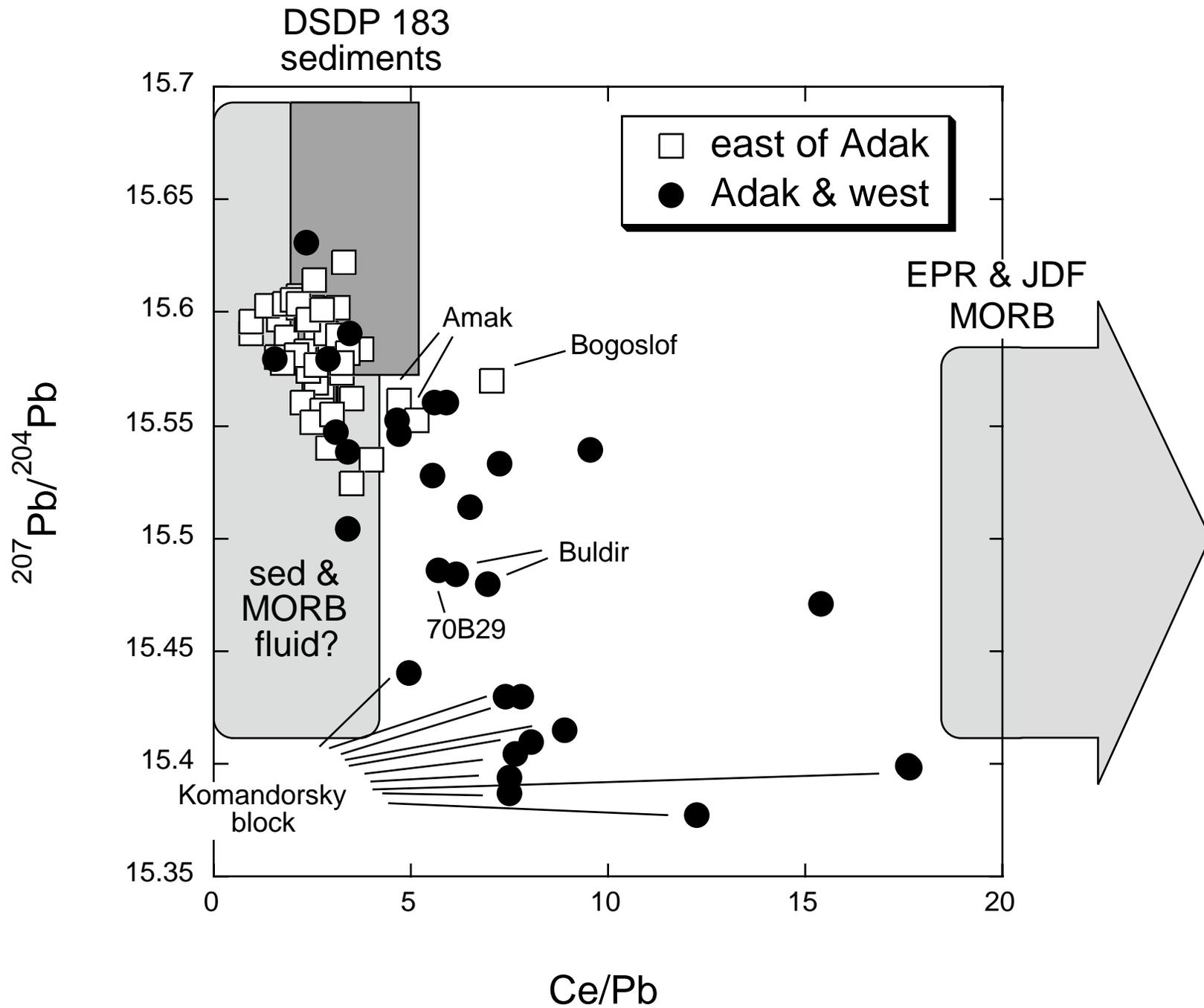


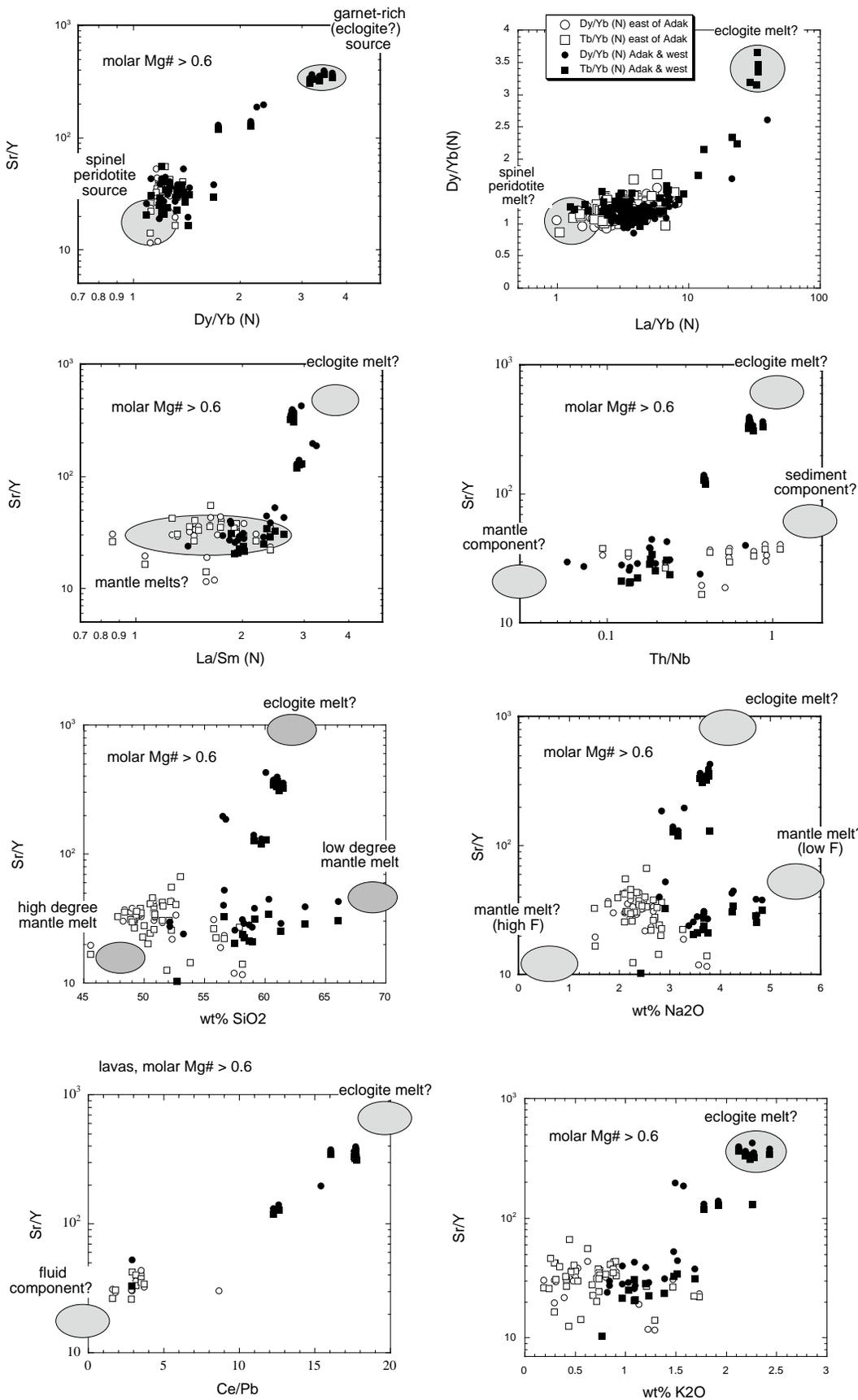




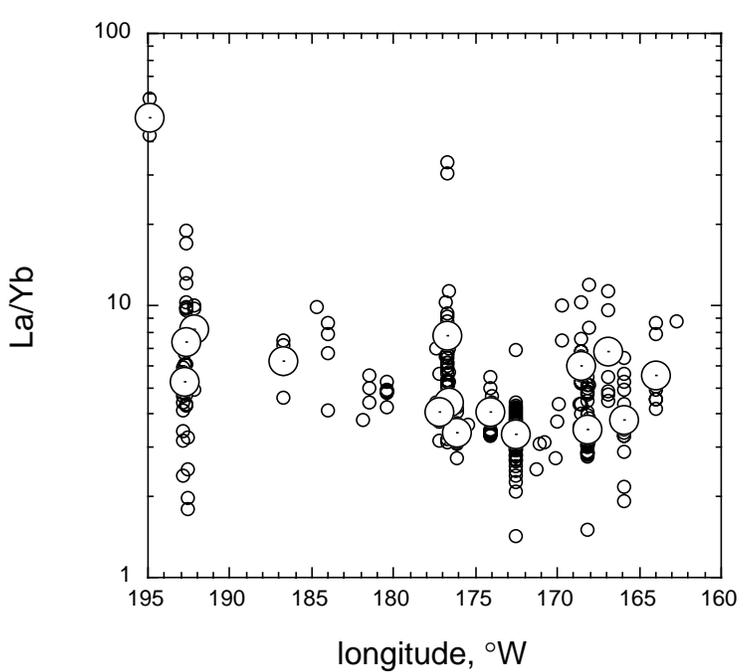
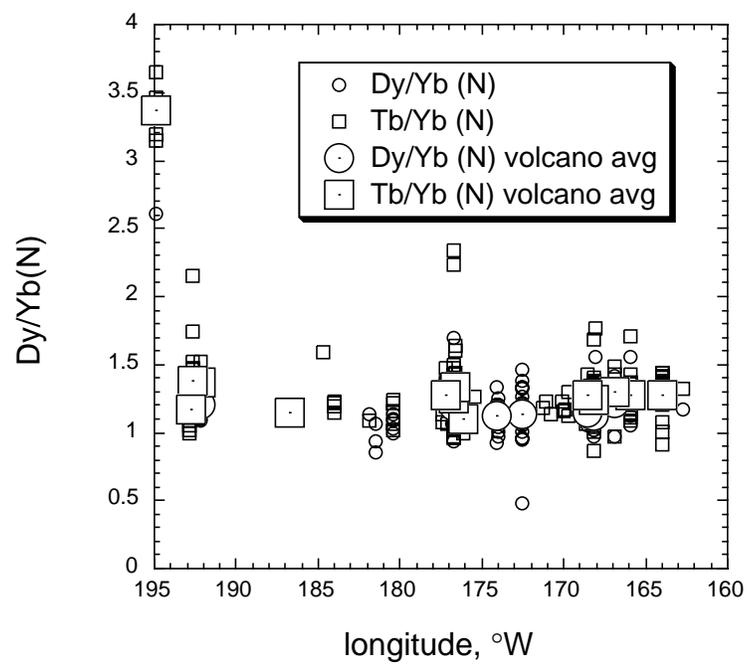
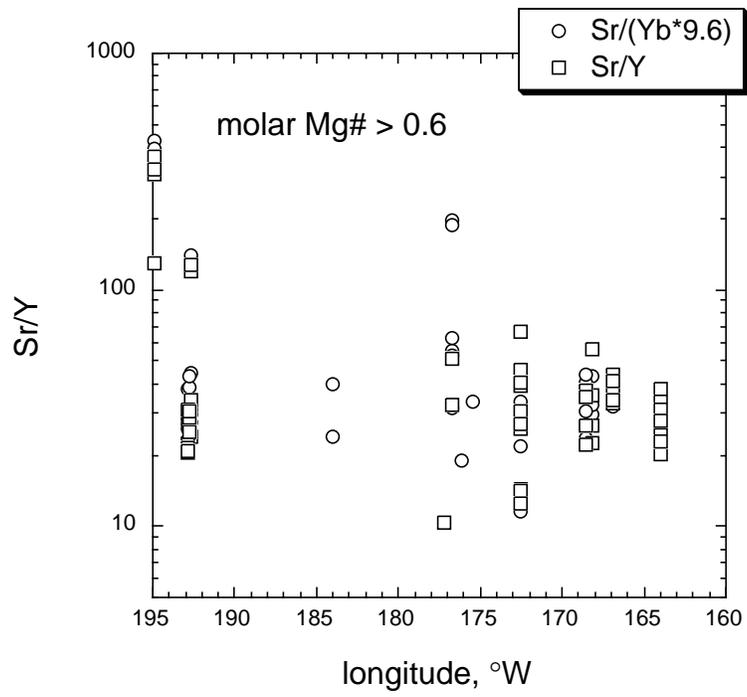


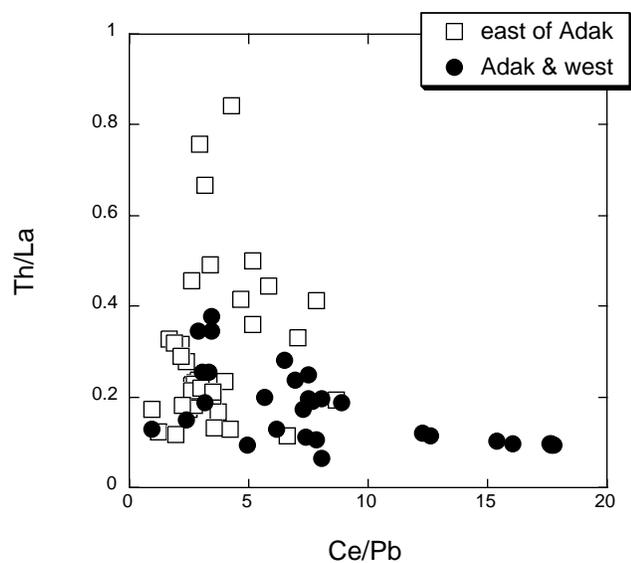
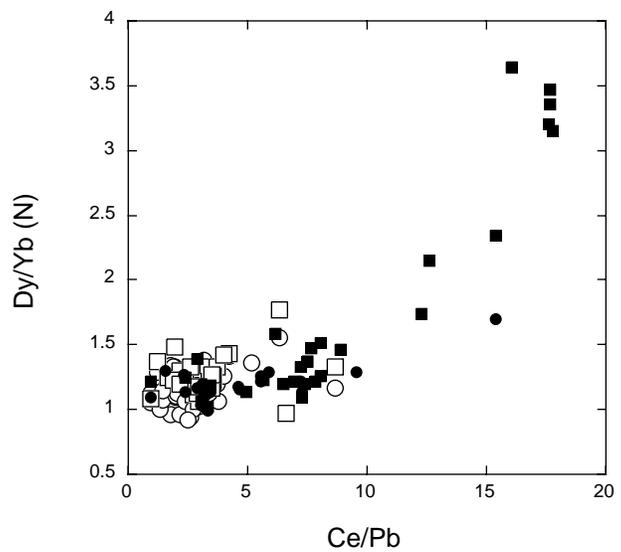
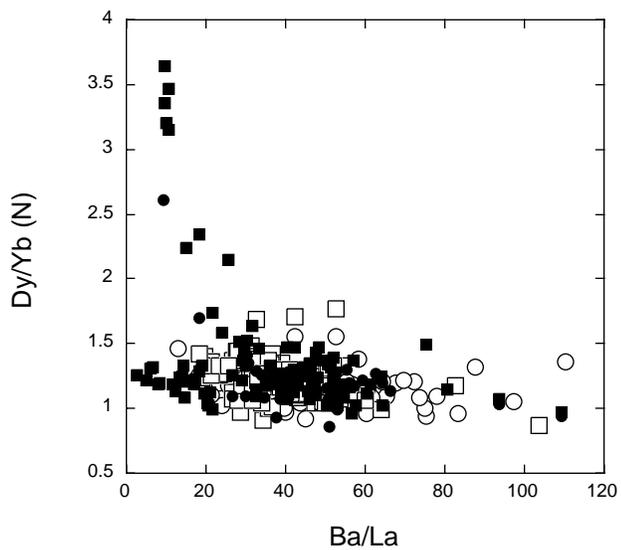
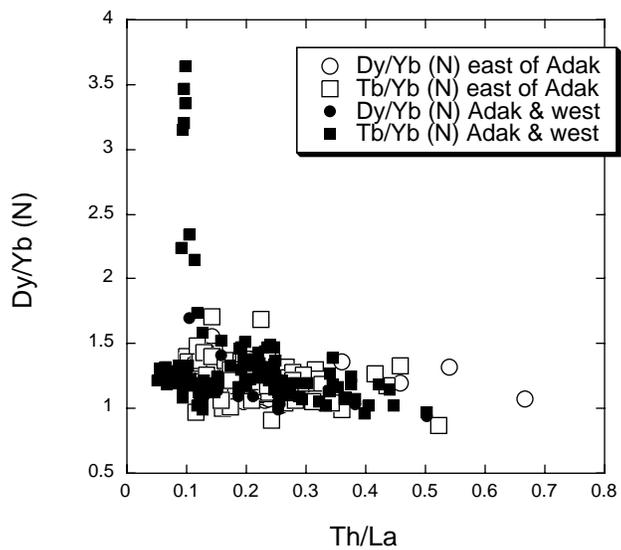




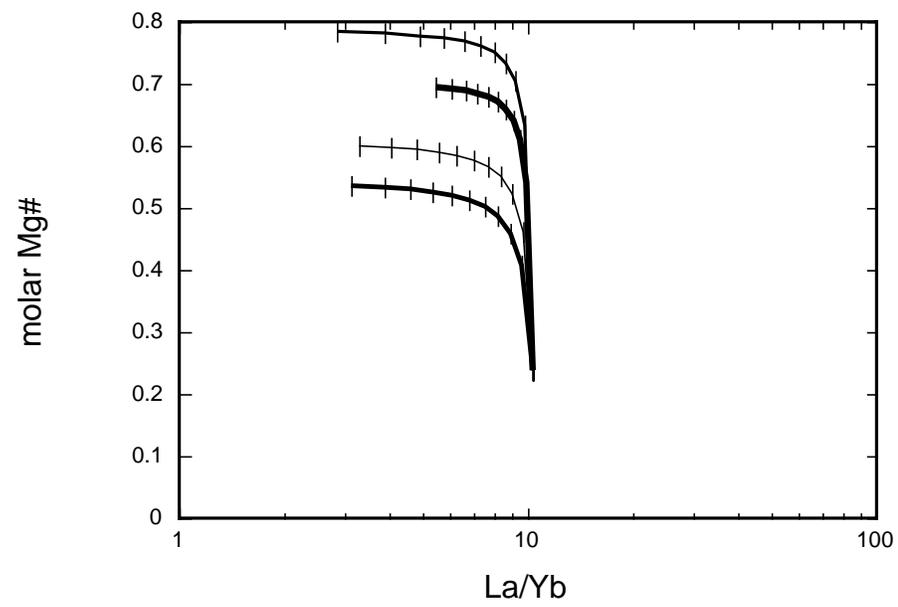
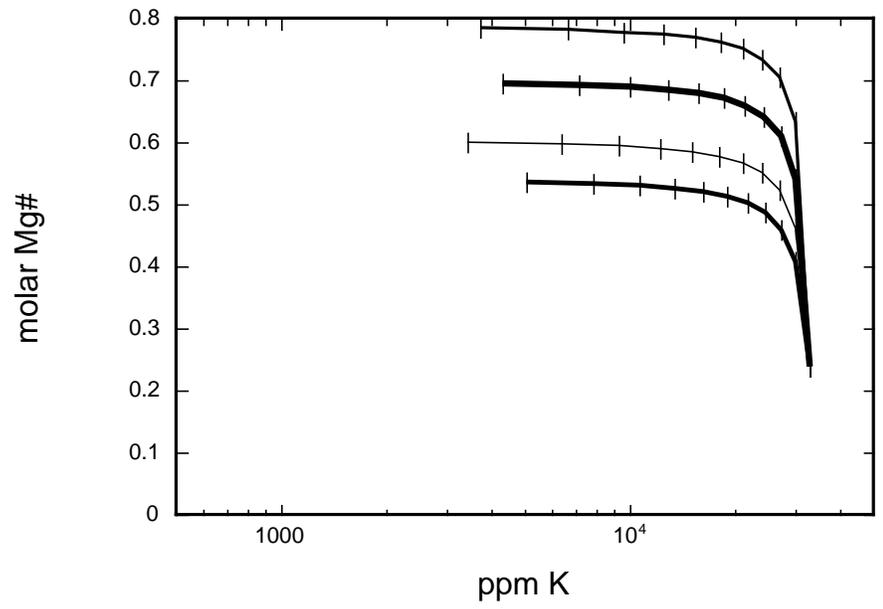
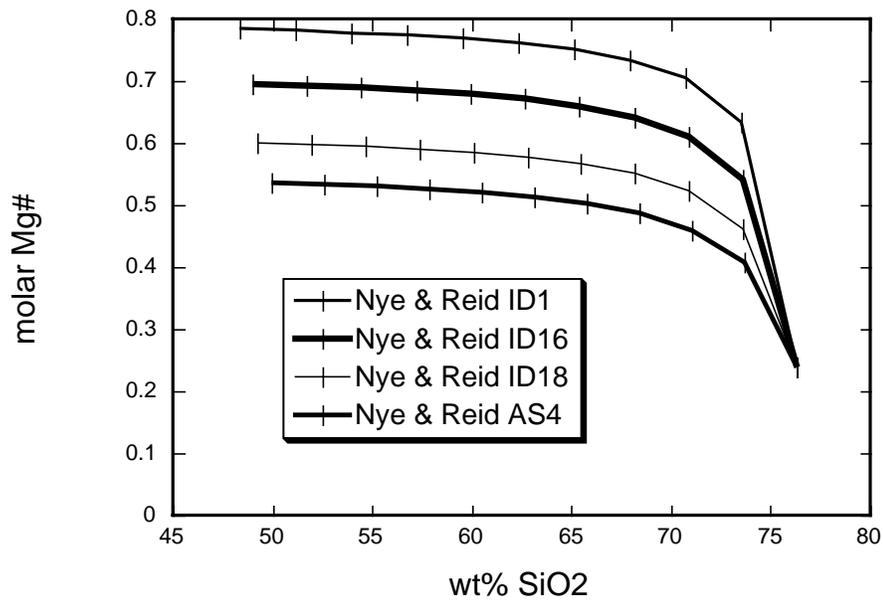


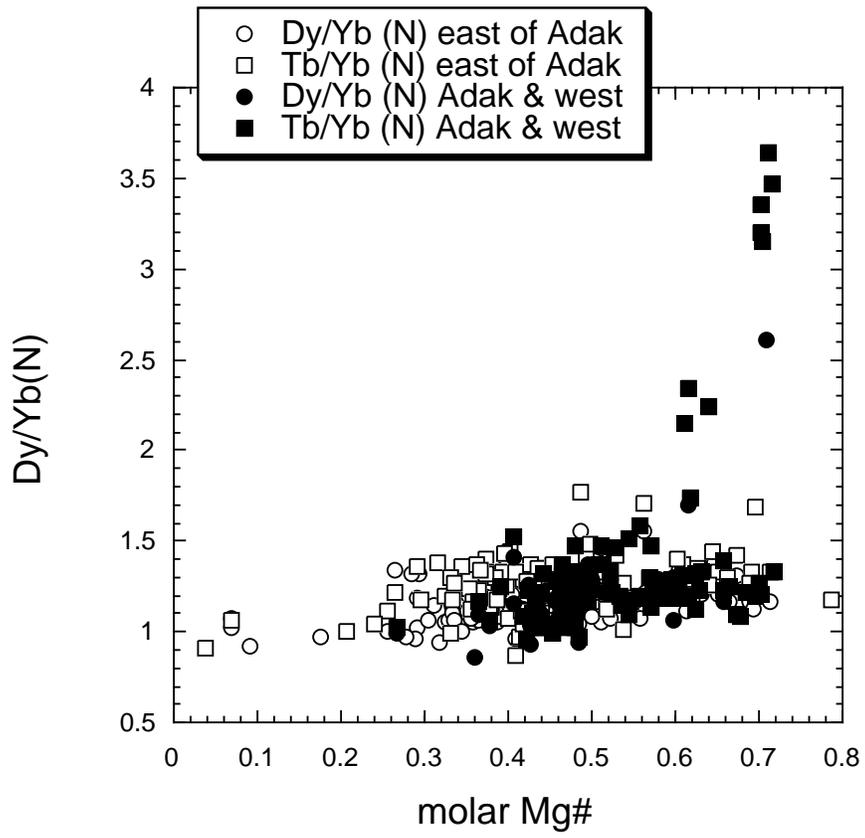
Kelemen et al., 2001, figure 16

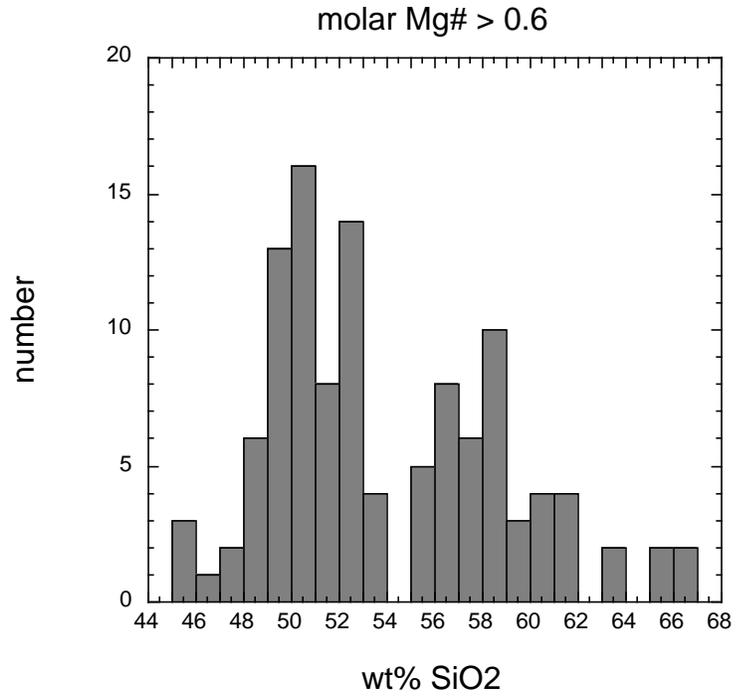


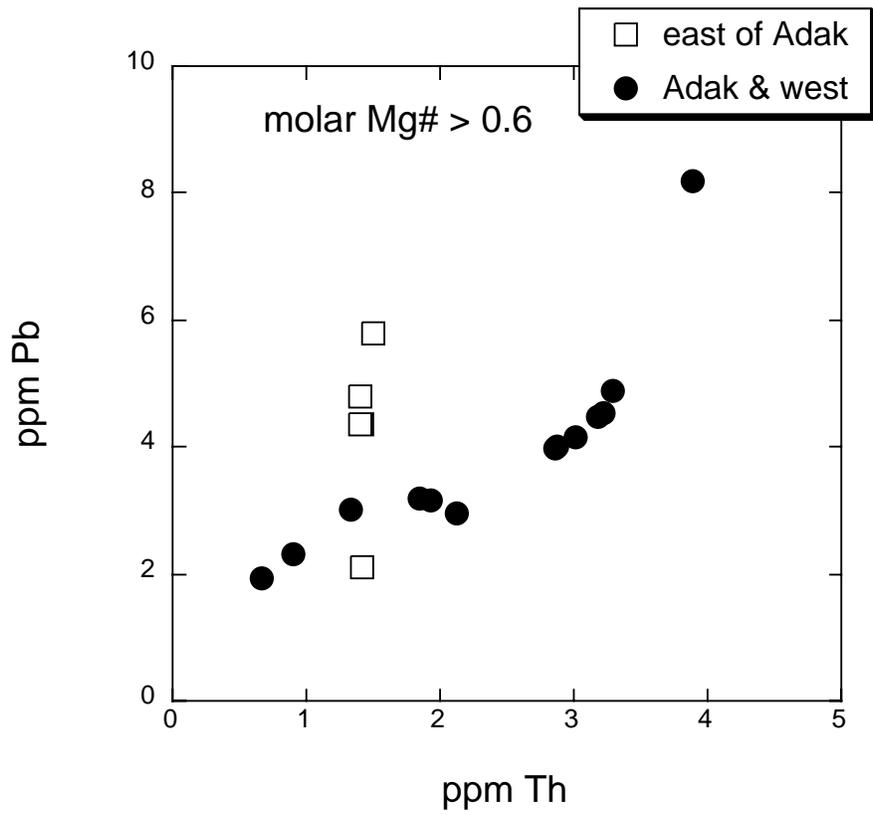


Kelemen et al., 2001, figure 18

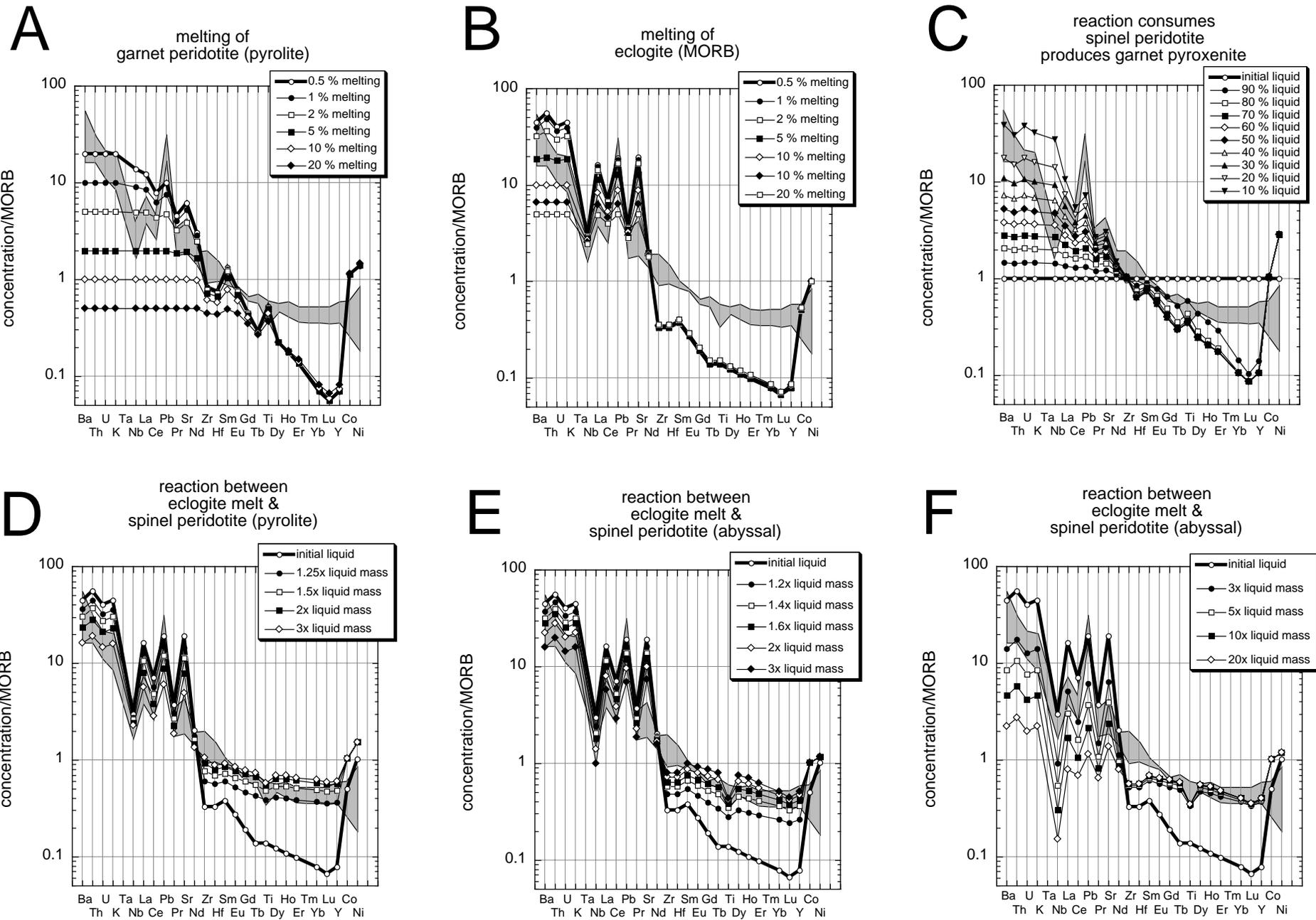


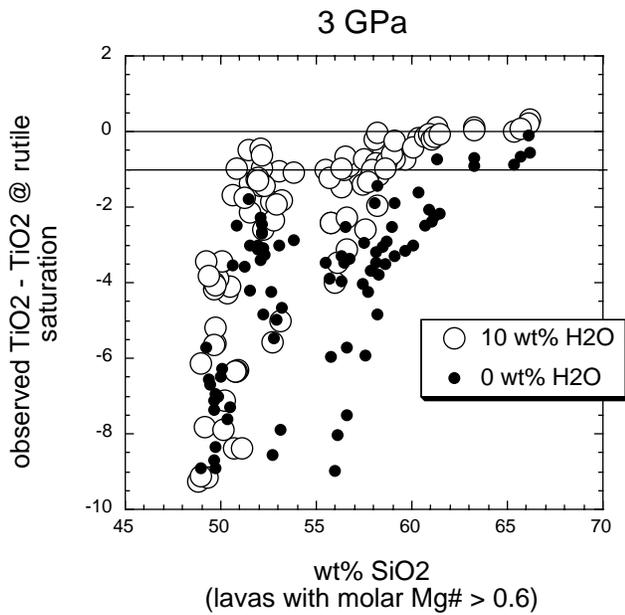
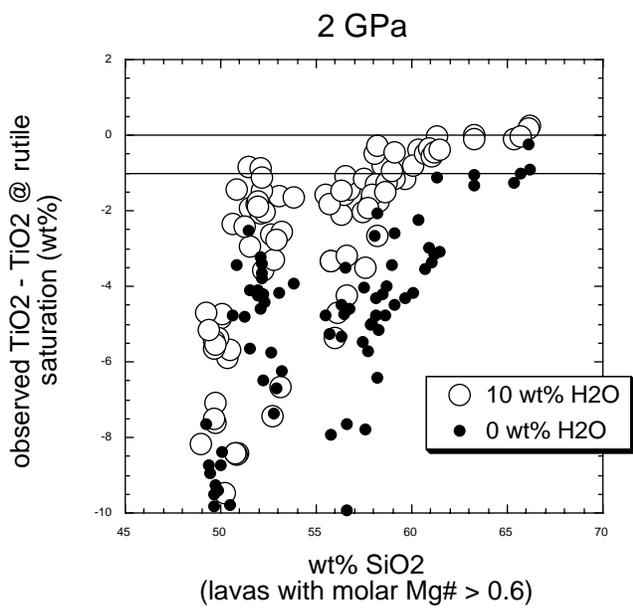
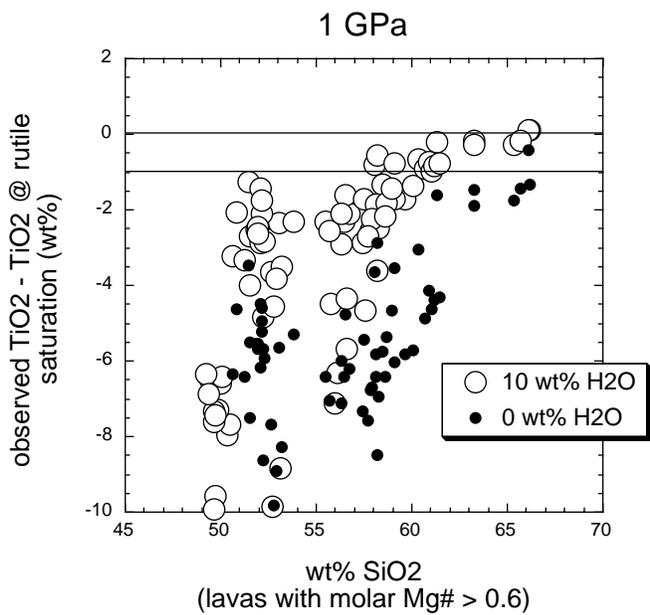






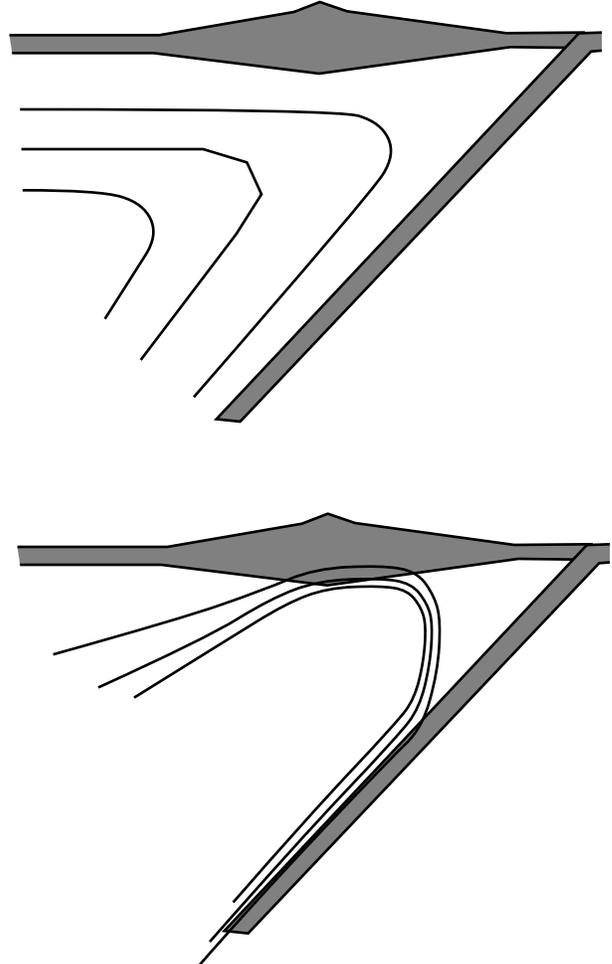
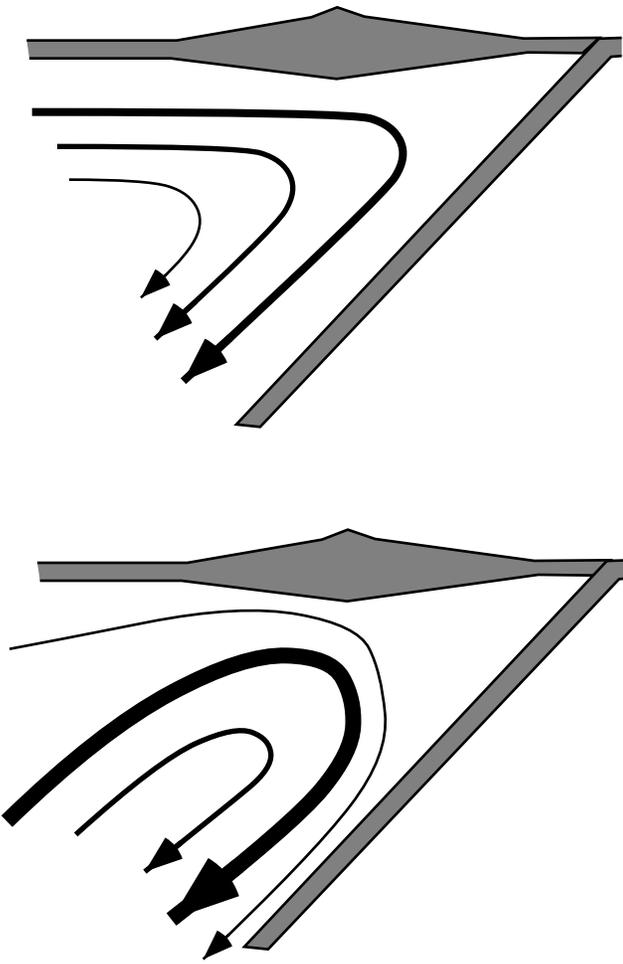
Kelemen et al., 2001, figure 22



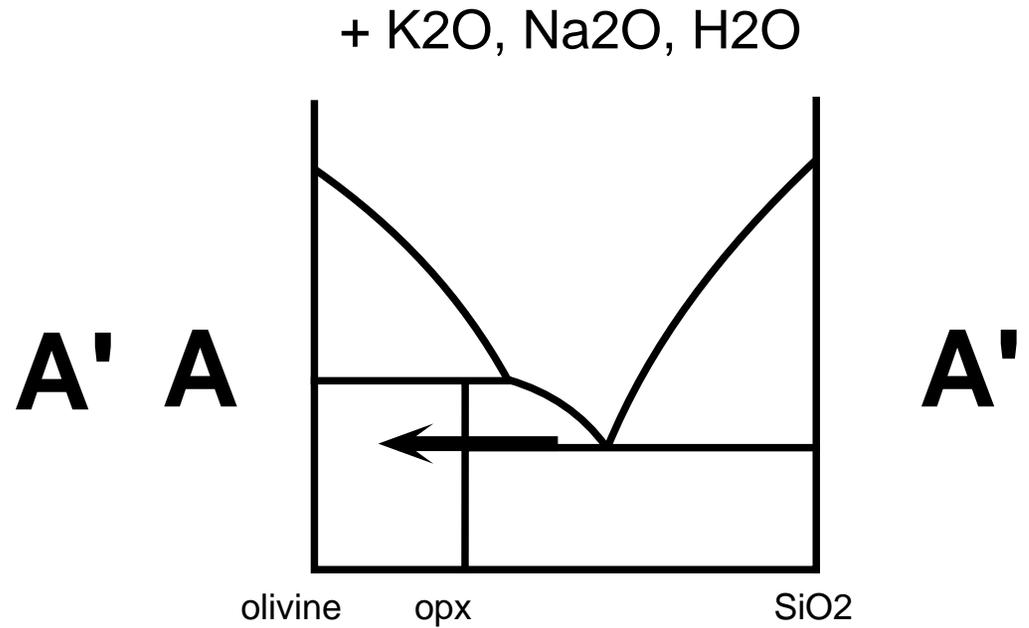
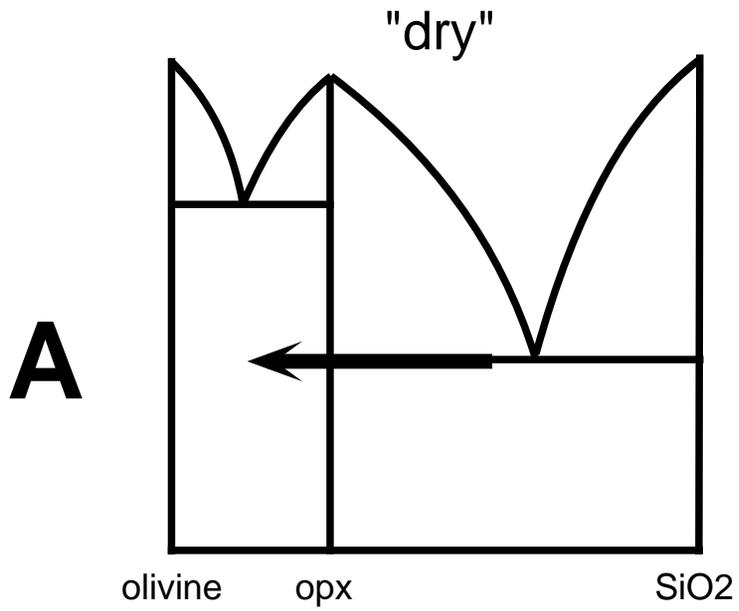
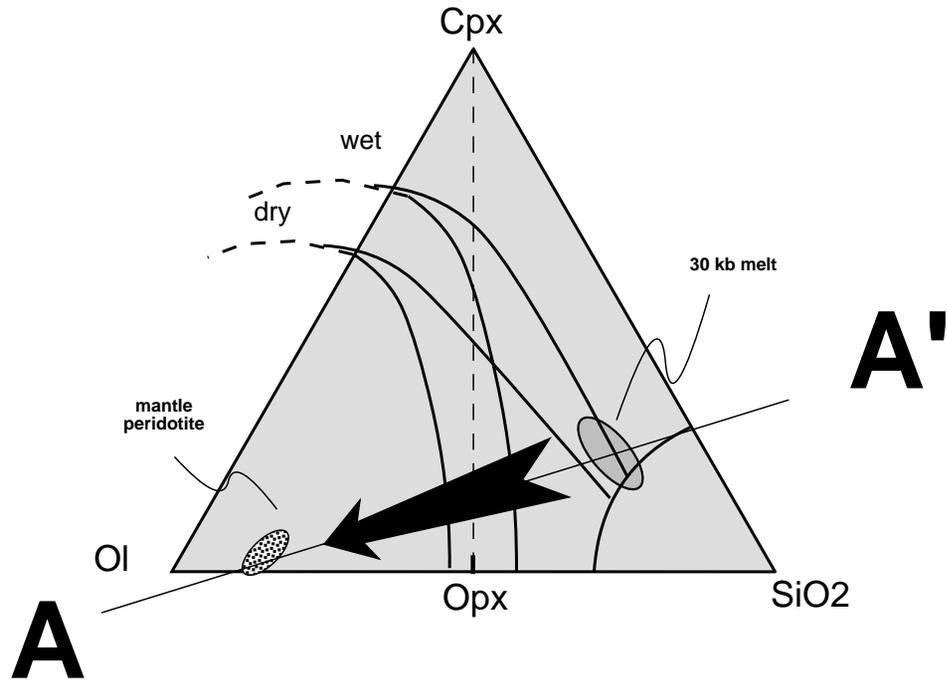


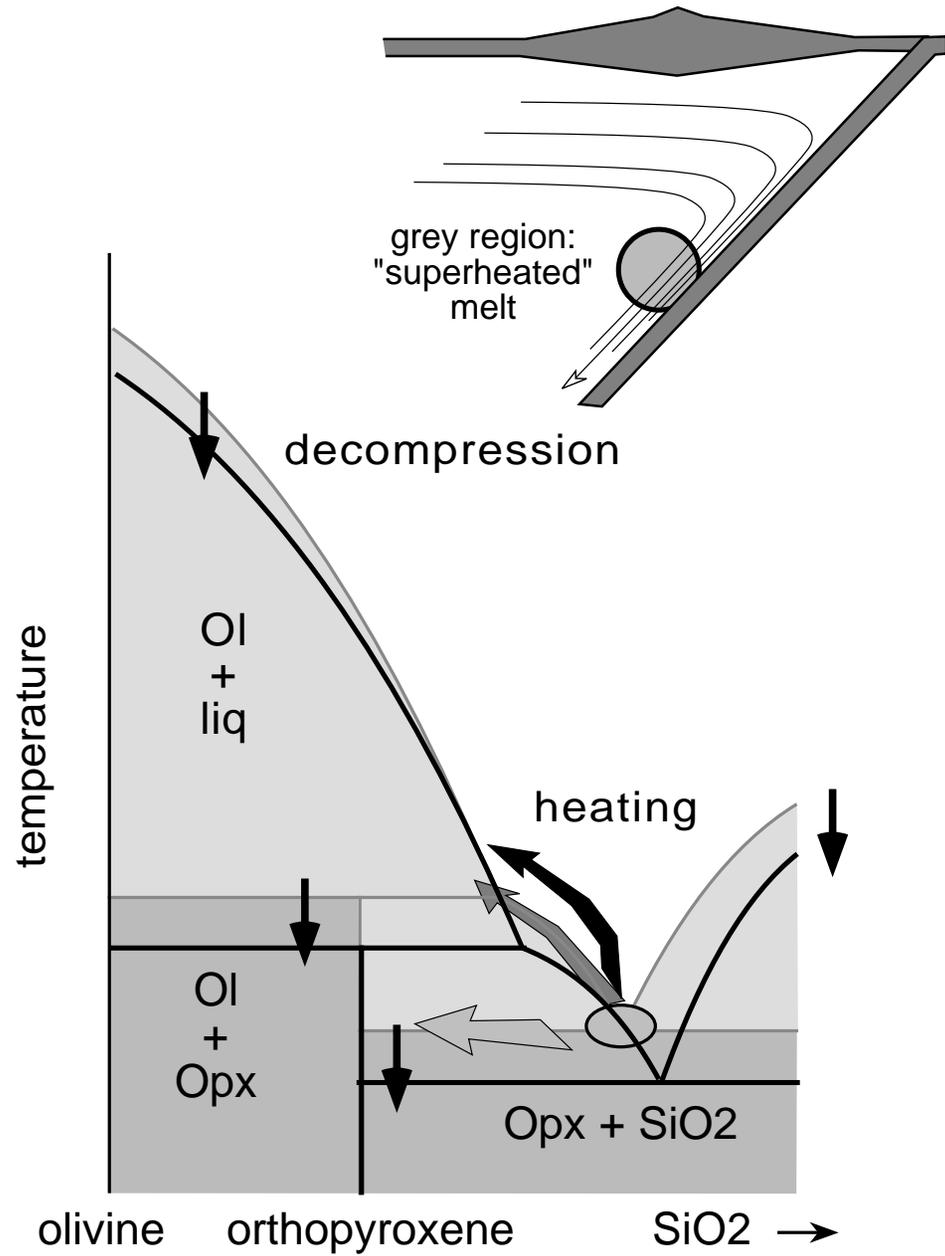
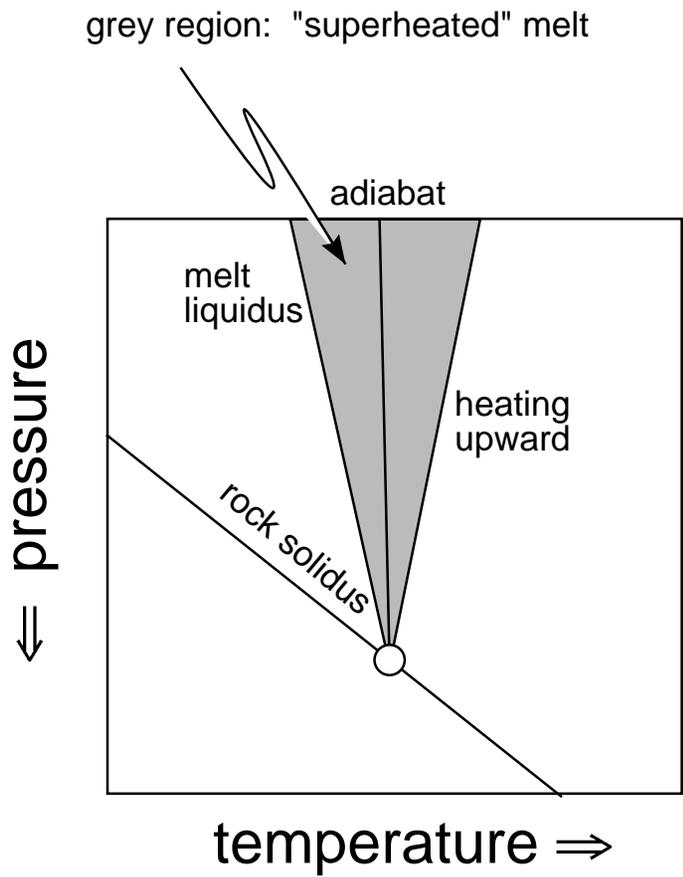
flow paths

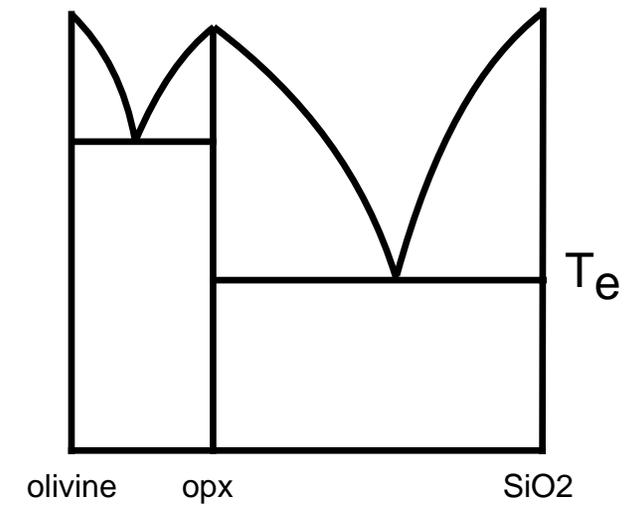
isotherms



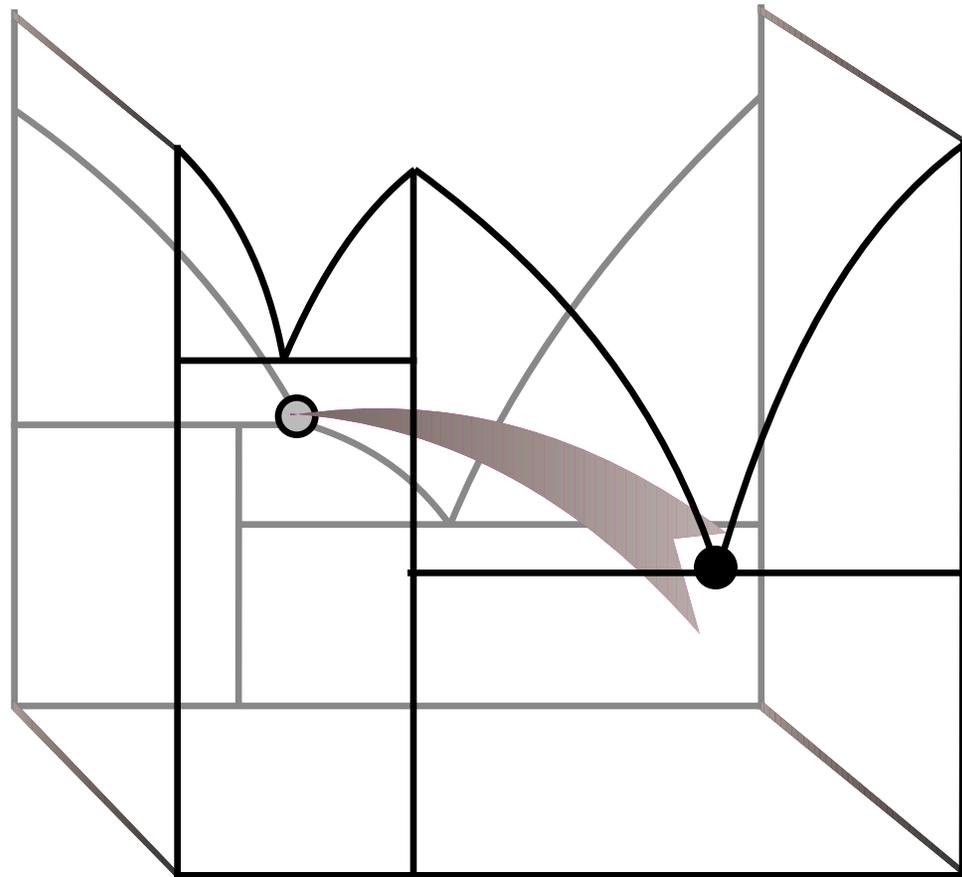
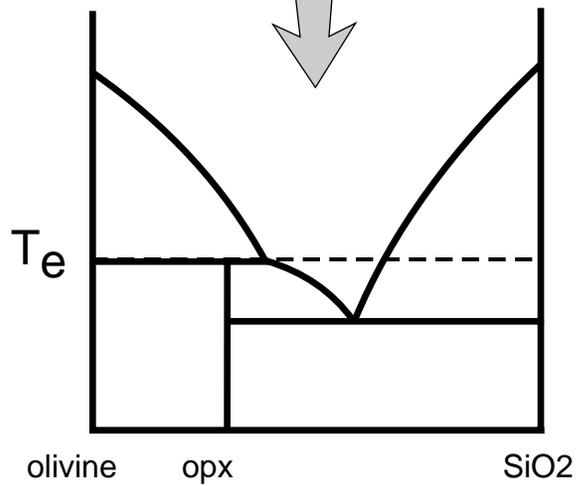
# thermal death



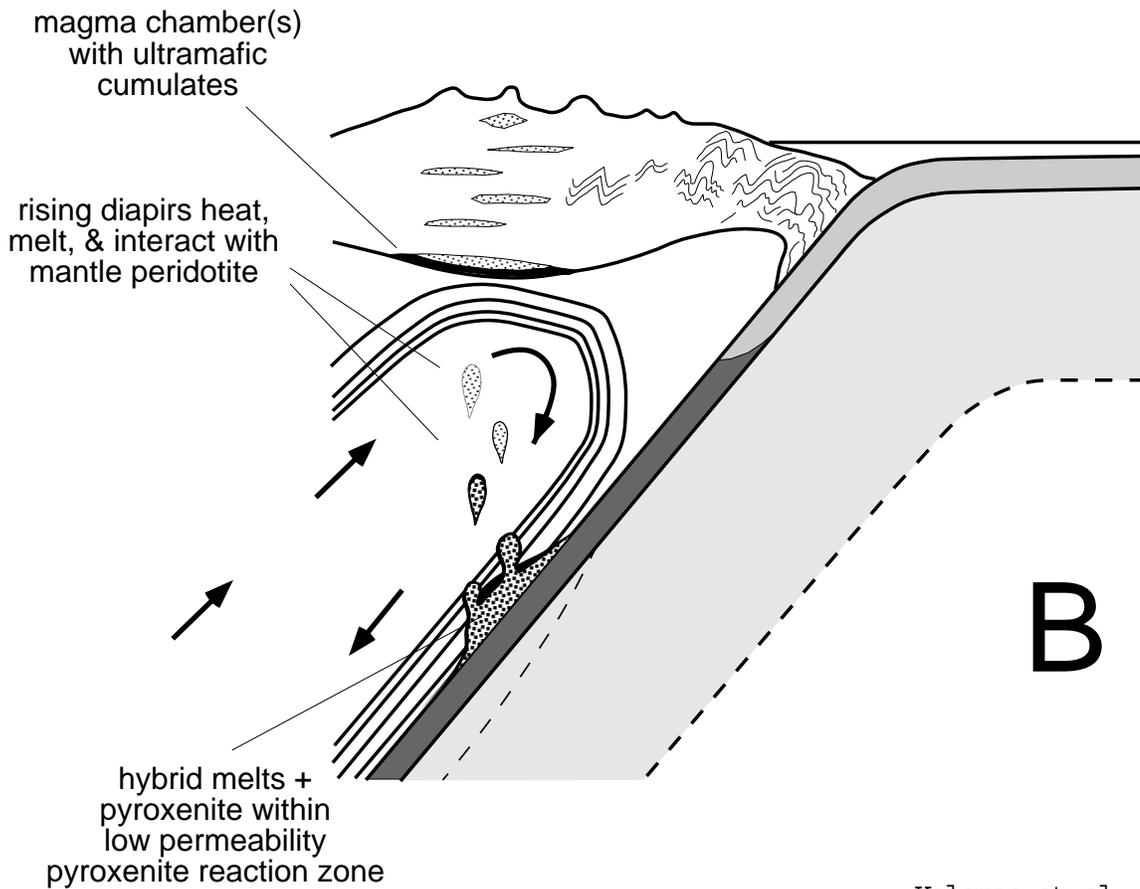
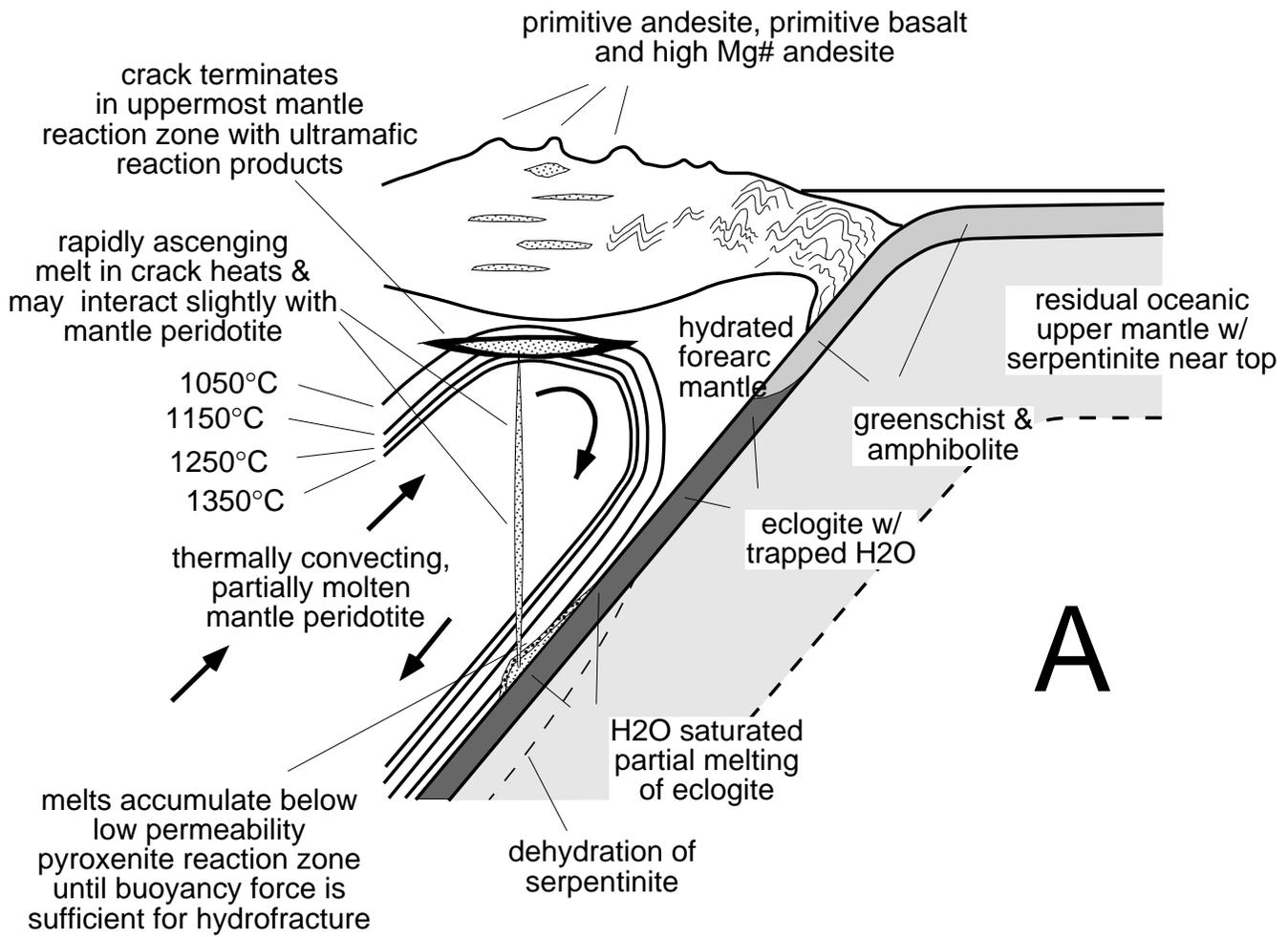




+ K<sub>2</sub>O, Na<sub>2</sub>O, H<sub>2</sub>O  
 due to decreasing  
 melt mass



isothermal reaction of "slab melt"  
 with mantle harzburgite



**Table 1: New Pb concentrations and isotope data for Aleutian lavas.**

lat (°N)	lon (°W)	location	sample	Ce, ppm *	Pb, ppm	206Pb/204Pb	207Pb/204Pb	208Pb/204Pb	isotope data from
52.370	184.020	Buldir	BUL6B	17.2	3.02	18.565	15.486	38.028	this paper
52.370	184.020	Buldir	BUL6A	20.5	3.48	18.641	15.514	38.132	this paper
52.370	184.020	Buldir	BUL4D	22.7	2.95	18.611	15.48	38.054	this paper
52.533	184.745	dredge near Buldir	70-B29	26.6	4.31	18.477	15.484	37.941	this paper
55.455	192.832	dredge, Piip volcano	V35G5B	15.0	1.92	18.089	15.431	37.535	Yogodzinski et al., 1994
55.425	192.732	dredge, Piip volcano	V35G1A	17.2	2.32	18.098	15.43	37.52	Yogodzinski et al., 1994
55.343	192.550	Komandorsky Islands	V35G8B	10.7	1.33	18.030	15.411	37.537	Yogodzinski et al., 1995
55.469	192.892	Komandorsky Islands	V35G7C	14.7	2.98	18.161	15.438	37.594	Yogodzinski et al., 1995

Pb concentrations determined by isotope dilution on hand picked rock chips leached in 2.5N HCl at 80°C for 1/2 hour, dissolved and analyzed on a Finnigan MAT Element I ICPMS. Estimated precision is  $\pm$  1% relative.

Pb isotope ratios determined on powdered, hand picked rock chips leached for 1h in 6.2N HCl at 100°C analyzed on a VG 354 mass spectrometer. Estimated precision, based on standard reproducibility, is  $\pm$  0.15% relative.

All results are corrected against NBS981 (Todd et al, 1996).

\*: Ce concentrations from Kay & Kay, 1994; Yogodzinski et al., 1994, 1995

**Table 2: New Sr and Nd isotope ratios for DSDP Hole 183 sediments**

sample	$^{143}\text{Nd}/^{144}\text{Nd}$	$\pm$ , ppm	$^{87}\text{Sr}/^{86}\text{Sr}$	$\pm$ , ppm	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$	Pb isotope data from
38-3, 20-21	0.512485	5	0.708201	7	19.093	15.645	38.814	Peucker-Ehrenbrink et al., 1994
39-1, 83-84	0.512510	5	0.708309	8	18.934	15.624	38.734	Peucker-Ehrenbrink et al., 1994
39-1, 108-109	0.512462	5	0.707979	7	18.546	15.596	38.546	Peucker-Ehrenbrink et al., 1994
39-1, 108-109	0.512416	6	0.707850	7				carbonate sample, no Pb data

Sr and Nd isotopic analysis were carried out with conventional techniques (Hauri and Hart, 1993). All results are corrected against LaJolla std  $^{143}\text{Nd}/^{144}\text{Nd}=0.511847$  and  $\text{NBS987}=\text{Sr}/^{86}\text{Sr}=0.710240$

\*\*\*\*\*

Table 3: Geodynamic setting of arcs with and without primitive Mg# andesites, based on references cited in this section, our compilation for the Aleutians, and Elliott et al. (this volume). Trace element (TE) evidence for subduction zone melting includes high Th/La and elevated Pb isotope ratios as outlined in Section 2.4.

	fast convergence (> 40 mm/yr)	slow convergence (< 40 mm/yr)
young subducting crust (< 40 Ma)	TE evidence for SZ melting no primitive andesites observed S Chile VZ Central America (except Panama) Central Andes†	TE evidence for SZ melting* primitive andesites observed* S Cascades Panama Chile Austral volcanic zone SW Japan
moderate age subducting crust (40 to 60 Ma)	TE evidence for SZ melting no primitive andesites observed CENTRAL ALEUTIANS	TE evidence for SZ melting** primitive andesites observed WESTERN ALEUTIANS
old subducting crust (>65 Ma)	TE evidence for SZ melting no primitive andesites observed Marianas Tonga Vanuatu	TE evidence for SZ melting** no primitive andesites observed Lesser Antilles

\*: Exception, southern part of South Sandwich arc

†: Exception, Puna Plateau

\*\* : Exception, northern part of South Sandwich arc

\*\*\*\*\*

**Table 4: Values used in trace element modeling.**

**Table 4A: peridotite melting, magma/mantle interaction**

element	D olivine/liq	refs	D opx/liq	refs	D cpx/liq	refs	D garnet/liq	refs	D spinel/liq	refs
olivine	1									
orthopyroxene			1							
clinopyroxene					1					
garnet							1			
spinel									1	
rutile										
Ba	1E-09	estimated ≈ D(K)	0.00001	estimated ≈ D(K)	0.00068	5	0.00001	estimated ≈ D(K)	0	estimated ≈ D(K)
Th	1E-09	estimated ≈ D(K)	0.00001	estimated ≈ D(K)	0.01	6	0.001	9	0	estimated ≈ D(K)
U	1E-09	estimated ≈ D(K)	0.00001	estimated ≈ D(K)	0.001	7	0.005	9	0	estimated ≈ D(K)
K	1E-09	1	0.00001	1	0.0072	1,5	0.00001	1	0	1
Nb	0.001	1	0.0029	1	0.0077	1,5	0.013	1	0.01	1
La	0.000007	1	0.001	1	0.0536	1,5	0.001	1	0.0006	1,12
Ce	0.00001	1	0.003	1	0.0858	1,5	0.008	1	0.0006	1,12
Pb	0.00001	estimated ≈ D(Ce)	0.003	estimated ≈ D(Ce)	0.072	1,5	0.0005	10	0	estimated ≈ D(K)
Pr	0.00004	interpolated	0.006	interpolated	0.13	interpolated	0.033	interpolated	0.0006	1,12
Sr	0.00001	1	0.003	1	0.1283	1,5	0.007	1	0	estimated ≈ D(K)
Nd	0.00007	1	0.009	1	0.1873	1,5	0.057	11	0.0006	1,12
Zr	0.004	1	0.04	1	0.1234	1,5	0.5	1	0.07	1
Hf	0.004	estimated ≈ D(Zr)	0.04	estimated ≈ D(Zr)	0.256	1,5	0.5	1	0.07	1
Sm	0.0007	1	0.02	1	0.291	1,5	0.217	11	0.0006	1,12
Eu	0.00095	1	0.03	1	0.33	1,5	0.45	11	0.0006	1,12
Gd	0.0012	interpolated	0.04	interpolated	0.37	interpolated	0.9	interpolated	0.0006	interpolated
Tb	0.0026	interpolated	0.05	interpolated	0.41	interpolated	1.5	interpolated	0.006	interpolated
Ti	0.015	1,2	0.15	1,2	0.4	1,2,5	0.6	1	0.15	1
Dy	0.004	1	0.06	1	0.442	1,5	2	11	0.0015	1,12
Ho	0.007	interpolated	0.065	interpolated	0.43	interpolated	2.8	interpolated	0.0023	interpolated
Er	0.009	1	0.09	1	0.43	8	3.5	11	0.003	1,12
Yb	0.023	1	0.1	1	0.43	1,5	7	11	0.0045	1,12
Lu	0.03	interpolated	0.12	interpolated	0.433	1,5	9	11	0.0053	1,12
Y	0.023	estimated ≈ D(Yb)	0.1	estimated ≈ D(Yb)	0.467	1,5	7	estimated ≈ D(Yb)	0.0045	estimated ≈ D(Yb)
Co	2	bulk D estimated ≈ 2	2	bulk D estimated ≈ 2	2	bulk D estimated ≈ 2	1	bulk D estimated ≈ 2	2	bulk D estimated ≈ 2
Ni	10	3	3.5	4	3	4	5	4	10	4

**Table 4: Values used in trace element modeling, continued**

**Table 4B: eclogite melting**

element	garnet D	refs	cpx D	refs	rutile D	refs
olivine						
orthopyroxene						
clinopyroxene			1			
garnet	1					
spinel						
rutile					1	
Ba	0.02	13	0.02	13,20	0	guessed
Th	0.001	estimated ≈ peridotite	0.03	21,22	0	guessed
U	0.005	estimated ≈ peridotite	0.04	21,22	0	guessed
K	0.02	13	0.02	13	0	guessed
Nb	0.05	14	0.02	21	30	25
La	0.08	15	0.04	23	0	guessed
Ce	0.2	13,15	0.08	23	0	guessed
Pb	0.005	estimated ≈ 10x perid	0.1	21	0	guessed
Pr	0.4	interpolated	0.14	interpolated	0	guessed
Sr	0.03	13,16,17	0.07	13,22,24	0	guessed
Nd	0.8	13,15	0.2	23	0	guessed
Zr	5	16	0.3	21,23	40	estimated > D(Nb)
Hf	5	15	0.3	20,23	40	estimated ≈ D(Zr)
Sm	5	13,15,16,18	0.4	23	0	guessed
Eu	7	13,15	0.45	13,21,22,23	0	guessed
Gd	10	13,15	0.6	23	0	guessed
Tb	14	13,15	0.65	interpolated	0	guessed
Ti	12	19	0.6	19	100	19
Dy	16	13,18	0.7	13,23	0	guessed
Ho	18	15,18	0.7	interpolated	0	guessed
Er	20	13	0.7	23	0	guessed
Yb	25	13,15,18	0.7	13,21,22,23	0	guessed
Lu	30	13,15	0.7	13,21,22,23	0	guessed
Y	25	16	0.8	21	0	guessed
Co	2	15	1	22	0	guessed
Ni	1	bulk D estimated ≈ 1	1	bulk D estimated ≈ 1	0	guessed



**Table 4: Values used in trace element modeling, continued**

**Notes**

1. Kelemen et al., 1993, references cited therein, and our unpublished ion probe data
2. Kelemen et al., 1990b
3. Arndt, 1977; Hart & Davis, 1978; Kinzler et al., 1990
4. Olivine/liquid from refs. in 3, plus mineral/olivine from Bodinier et al., 1997; Kelemen et al., 1998
5. Hart & Dunn, 1993
6. Value based on Hauri et al., 1994 and LaTourrette & Burnett 1992; for our purposes, these are sufficiently close to other values suggested by Beattie, 1993; Lundstrom et al., 1994; Salters & Longhi, 1999; Wood et al., 1999
7. Value based on Beattie, 1993; for our purposes, these are sufficiently close to other values suggested by Hauri et al., 1994; LaTourrette & Burnett 1992; Lundstrom et al., 1994; Salters & Longhi, 1999; Turner et al., 2000
8. Adjusted from value in Hart & Dunn (1993) to give smooth REE pattern for garnet peridotite and eclogite melting.
9. Values based on LaTourrette & Burnett, 1992; Beattie, 1993; and Hauri et al., 1994; for our purposes these are sufficiently close to other values suggested by Salters & Longhi, 1999
10. Hauri et al., 1994; Beattie, 1993; Salters & Longhi, 2001
11. Shimizu & Kushiro, 1975
12. Stosch, 1982
13. Philpotts & Schnetzler 1970a,b; Schnetzler & Philpotts, 1970
14. Interpolated based on garnet D pattern in ref. 1 and K, La values
15. Irving & Frey, 1978
16. Green et al., 1989
17. Jenner et al., 1994
18. Nicholls & Harris, 1980
19. Ti is a major element during partial melting with residual rutile, and its concentration is controlled by phase equilibrium rather than partitioning. However, for modeling, Ti value is interpolated based on garnet and cpx D patterns in refs. 1, 5, plus and Tb, Dy values in this Table, then adjusted to fit observation by Rapp et al., 1999, that Ti is not fractionated from REE during rutile saturated partial melting of eclogite.
20. Luhr & Carmichael, 1984
21. Larsen, 1979
22. Dostal et al., 1983
23. Fujimaki et al., 1984
24. Hart & Brooks, 1974
25. Green & Pearson, 1987
26. Rapp papers
27. Kelemen et al., 1992
28. Dick, 1989
29. Based on ~5% cpx (exsolved from high temperature opx) and cpx data from Johnson et al., 1990; Johnson & Dick, 1992