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R. W. Kay

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Elemental abundances relevant to identification of magma sources

By R. W. Kay

Department of Geological Sciences Cornell University, Ithaca, New York 14853, *U.S.A.*

The search for chemical characteristics of magma sources is usually done by analysing the magmas themselves. This indirect approach has limitations: clearly the magma has only some of the source's characteristics. What we require are process-independent chemical characteristics, analogous to the isotopic abundance of radiogenic daughter isotopes that have been used so successfully in defining magma sources. Process-independent chemical characteristics in mid-oceanic ridge, oceanic island and island-arc basalts (m.o.r.b., o.i.b., i.a.b.) have been used to identify contrasting chemical characteristics of mantle peridotite from these three tectonically distinct regions. As an example, the abundance ratios of one group of elements (e.g. Cs, K, Rb, Ba, U, and perhaps Th) relative to another group (e.g. light r.e.e., Zr, Hf) are found to be fractionation-independent during most shallow-level basalt fractionation. These ratios are presumed to reflect the chemical characteristics of the mantle source of basalt from the three tectonic environments. In particular the ratios indicate the large cation-depleted nature of all m.o.r.b. and most o.i.b. peridotite sources. In common with many other island arcs, the abundance ratios are consistently higher in mantle under the Aleutian arc than in adjacent non-arc mantle represented by oceanic ridge, oceanic island, and back-arc basalts. The contention that subduction of sediment could result in arc mantle sources with these high ratios is substantiated by trace element analyses of Ba and Cs-rich deep sea sediments of the type that are being subducted at present at the Aleutian trench. The importance of recycling of sediment into the mantle at island arcs as an important control on the trace element (and isotopic) evolution of the mantle is indicated.

Introduction

The compositionally distinct regions of the upper several hundred kilometres of the Earth are: upper crust, lower crust and upper mantle. Each of these regions serves as a magma source or at least as a magma component. The major element compositions of these contrasting regions are summarized in table 1. Mineralogically, the mantle is dominantly peridotite and the crust feldspathic but compositionally diverse igneous, sedimentary and metamorphic rocks. Mainly we know this structure through geophysical methods, through analysis of magmas, and to some extent by examination of samples from these distinct regions, both those exposed on the Earth's surface and those brought up as xenoliths in lavas and kimberlites (see, for example, Kay & Kay 1981).

The subject of this paper is magma sources, the solid rocks that yield magmas. Igneous differentiation (crystal settling, magma mixing and other processes affecting magmas) tells us nothing about magma sources; it gets in the way of the discussion. Yet, if differentiation-independent characteristics of the magmas are identified, characteristics of 'primary' magmas can be identified, and their characteristics do tell us about the source.

Radiogenic isotopic compositions are differentiation-independent characteristics of magmas (see, for example, Allègre 1982; Allègre *et al.* 1980; Anderson 1982; Stosch *et al.* 1980; White & Hofmann 1982), but will hardly be mentioned here: the topic here is elemental composition.

Table 1. Crustal and mantle compositions (percentages by mass†)

	lower crust 'petrological models'			whole crust	upper crust	mantle
	1	2	3	4	5	6
SiO_2	54.0	41.02	61.2	62.5	65.2	38.3
TiO_2	0.9	1.50	0.54	0.68	0.57	0.22
Al_2O_3	19.0	17.86	15.6	15.6	15.8	3.97
Fe_2O_3	—	—	3.07	6.1	1.2	—
FeO	9.0	13.26	2.57	—	3.4	7.8
MnO	—	0.12	0.08	0.13	0.09	0.16
MgO	4.1	9.24	3.36	3.2	2.2	38.3
CaO	9.5	11.72	5.57	6.0	3.3	3.5
Na_2O	3.35	1.95	4.42	3.4	3.7	0.37
K_2O	0.6	0.41	1.03	2.3	3.23	—
U	0.2	—	0.05	—	2.45	—
Th	0	—	1	—	10.3	—
Ba	433	—	779	—	1070	—

(1) Taylor's (1977) island arc model: lower crust as residue.
(2) Hornblende granulite, Ichinomegata Japan.
(3) Sheraton *et al.*'s (1973) Scourian gneisses, Scotland.
(4) Holland & Lambert's (1972) continental crust (highly weighted toward granulite facies rocks).
(5) Crystalline rocks from Canadian Shield (Eade & Fahrig 1968).
(6) Sun (1983).

† U, Th, Ba in micrograms per gram.

This could be regarded as a fundamental liability, since elements of variable isotopic composition are in effect 'colour-coded' and in practice, combined discussion of isotopes and trace elements yield more satisfactory conclusions than either does alone.

While focusing on the chemical characteristics of magma sources, the scope of discussion will include processes that have created or modified the source, especially processes at convergent plate margins.

Most of the compositional data relevant to magma sources concerns mantle sources (peridotite). The plan of this paper is to review mantle sources in the three tectonic environments where basalts occur: divergent plate margins, convergent plate margins and intraplate regions. The compositional diversity of potential magma sources that involve sedimentary rocks will then be illustrated by discussion of specific examples from the Aleutian island arc. Crustal magma sources will then be considered.

Peridotite source of mid-oceanic ridge basalt (m.o.r.b.)

Some compositional features of basalt magmas reflect the composition of their peridotite sources. This is convenient, because direct analysis of mantle samples from all tectonic régimes is limited by inefficient and biased sampling: for instance the tholeiitic basalts lack mantle xenoliths. Thus, our main reason for believing that the mantle is heterogeneous is that magmas coming from the mantle are heterogeneous in a number of fractionation-independent characteristics (see, for example, O'Nions *et al.* 1980). It is probably worthwhile retracing the arguments relevant to mid-ocean ridge basalt, m.o.r.b.

(1) There seems little argument that most of the m.o.r.b. lavas that are *not* at equilibrium with Fo_{90-92} olivine are modified by either crystal–liquid fractionation or magma mixing or both. This includes most m.o.r.b. magmas.

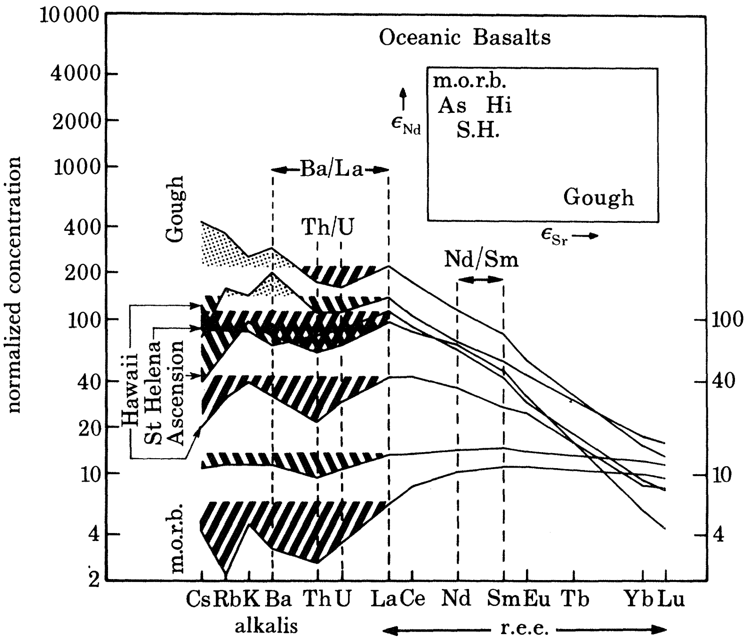


Figure 1. Concentrations of some trace elements in a variety of oceanic basalts plotted as multiples of chondritic concentration (for r.e.e., Th, U and Ba), and (for Cs, Rb and K) as multiples of values that yield the same normalized values in m.o.r.b. KD11 (for Cs, Rb, K and Ba). Gough Island sample is plotted at twice its actual level for clarity.

Elements are arranged in order of increasing incompatibility (in a Ca lattice site) from right to left. Note that the normalized concentration ratios of the largest (and most 'incompatible') trace cations (Cs, Rb, K, Ba, Th and U) to the most incompatible r.e.e. (La) are less than one (striped pattern) for m.o.r.b. For some oceanic island basalts the ratios are also often less than one, but for some samples are greater than one (solid pattern). The ratios for the Gough island sample are highest. The variability is thought to mirror variability in the mantle. The inset shows that isotopic systems (Nd and Sr are shown, after Zindler *et al.* 1982) are also variable, indicating that long-lived mantle heterogeneities are represented. Gough island has the highest parent–daughter ratio for the Sr isotope system and the lowest for the Nd isotope system, correlating with parent–daughter ratios observed in the basalts themselves.

(2) The m.o.r.b. source has low ratios of large to small cations (see figure 1 and table 2). Based on the minerals that control observed magmatic trends, and on knowledge of crystal-melt partition coefficients there is no way to decrease large cation to small cation ratios in a m.o.r.b. differentiation series. These elements should be enriched in successive residual magmas without fractionation from each other. Trace element systematics that provide a test of these criteria are widely recognized and applied (see Langmuir & Hanson 1980; Cox & Clifford 1982; le Roux *et al.* 1981; Tarney *et al.* 1980). Uranium-series disequilibrium studies on recent volcanic rocks (Allègre & Condomines 1982; Newman *et al.* 1982) provide an additional test of trace element fractionation.

(3) Detailed description of processes of magma modification are irrelevant to the question of source composition, provided that process-independent characteristics are being considered. Characteristics such as the Ni content of the source or the Ca/Al ratio of the source are difficult to infer from magma composition for they are not process independent (see Dawson *et al.* 1980). Ratios of incompatible elements are process-independent, and we can determine which elements are incompatible (see point 2 above).

For these reasons, it is easy to see why it is commonly held that variations in large cation ratios (not concentrations) reflect variations in the mantle. I do not propose to review the

Table 2. Trace elements in volcanic rocks from mid-ocean ridges, oceanic islands and island arcs†

	1	2	3	4	5	6	7	8	9	10
Ba	1104	128	360	12.4	46	132	925	630	291	336
La	7.85	3.81	11.2	2.3	4.9	14.4	48.9	39.7	38.9	35.2
Ce	20.0	9.9	27.3	7.8	12.7	37.5	96.2	80.7	80.1	79.1
Nd	15.1	6.3	18.8	6.9	9.7	24.1	43.9	37.5	39.6	41.6
Sm	3.76	1.77	5.05	2.4	3.08	6.16	9.37	7.95	8.3	10.7
Eu	1.45	0.600	1.59	0.94	1.13	2.06	2.90	2.53	2.63	3.55
Tb	—	0.37	0.91	—	—	1.09	1.08	0.92	0.98	1.64
Yb	1.53	1.26	3.15	2.3	2.81	1.89	1.30	1.80	2.09	4.03
Lu	—	0.200	0.47	0.34	0.408	0.277	0.176	0.242	0.284	0.589
Cs	0.82	0.33	0.97	0.019	0.048	0.073	0.34	0.90	0.36	0.14
Th	3.34	0.50	2.2	0.13	0.454	1.10	4.72	3.59	3.22	2.53
U	0.82	0.73	1.07	0.050	0.152	0.41	1.70	1.15	1.19	0.96
Hf	3.54	1.27	3.4	—	—	4.77	5.22	6.87	6.47	6.84
Ta	—	0.073	0.60	—	—	1.00	5.8	2.16	3.28	2.99

(1) i.a.b. Bogoslof Is. B1927 basalt.
(2) i.a.b. Umnak Is. 45–95 tholeiitic basalt.
(3) i.a.b. SAR-11 Westdahl Volcano, Aleutians, tholeiitic basalt.
(4) m.o.r.b.-D1 tholeiite.
(5) m.o.r.b.-KD11 tholeiite.
(6) o.i.b. Hawaii BHVO-1 tholeiite.
(7) o.i.b. Hawaii K17 nephelinite.
(8) o.i.b. Gough G111 alkali basalt.
(9) o.i.b. St Helena SH 2878 alkali basalt.
(10) o.i.b. Ascension ASC2765 alkali basalt.

† References: McCulloch & Perfit (1981); Kay & Hubbard (1978); Oversby (1969). New analyses by instrumental neutron activation (see Kay *et al.* 1983). B1927 is sample of Arculus *et al.* (1977).

variability itself or the constancy of some large cation ratios (see Kay & Hubbard 1978; Bougault *et al.* 1980; Hofmann & White 1983): perhaps exemplified by Iceland–Reykjanes Ridge trend. Note that the mixture of peridotite sources as opposed to basalt magmas can be determined from trace element systematics of highly compatible elements (see, for example, Langmuir *et al.* 1977).

Peridotite sources of oceanic island basalts o.i.b.

To a variable extent the trace element ratios of o.i.b.-type peridotite are not reflected in o.i.b. basalts. Figure 1 shows that the concentrations of a number of large cations are much higher in oceanic island basalts than in m.o.r.b., but that smaller cations (e.g. heavy r.e.e.) are similar in concentration level. For the entire range of elements in figure 1, ratios of large to small cations has increased, perhaps substantially, by melting processes that yield o.i.b. Were it not for isotopic distinctness, we would have a lot of argument over whether m.o.r.b. sources and o.i.b. sources are the same (see Cohen & O'Nions 1982; Allègre & Condomines 1982), because as the percentage of melting decreases, the role of accessory minerals in fractionating the trace elements in figure 1 becomes an increasingly plausable alternative to source heterogeneity as an explanation for large cation heterogeneity of the magmas. Generally, isotopic data are required to discriminate between the alternatives-trace element systematics are not sufficient.

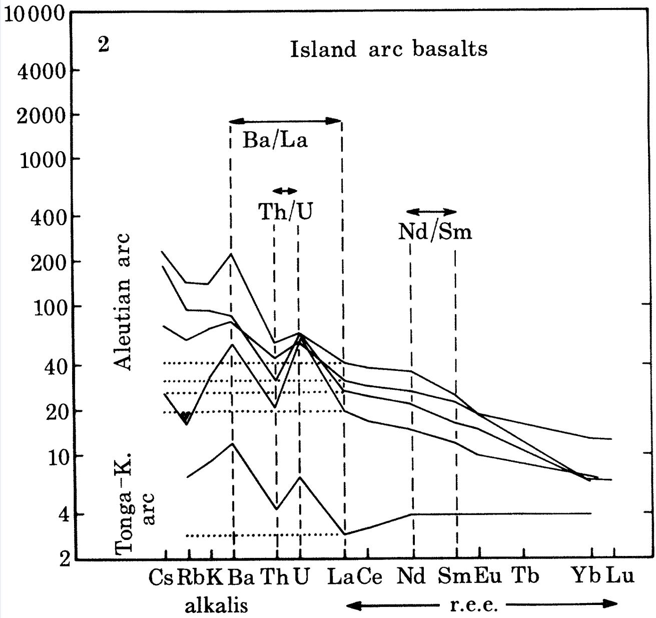
Still, as in m.o.r.b., for elements with very low solid–liquid distribution coefficients (and this probably includes all elements to the left of La in figure 1), ratios in melt and mantle are nearly identical (see, for example, Clague & Frey 1982; Hofmann & White 1983). If this is true, then the variations in ratios such as Ba/Cs, K/Rb and Ba/La in the o.i.b. population that are depicted in figure 1 are mantle variations. The heterogeneity in isotope systems in the same oceanic islands is well known (see figure 1) and implies that the heterogeneities are long-lived, some approaching the age of the Earth (Allègre *et al.* 1983). Their scale and location is the subject of much debate. Note that equilibrium fractionation between crystals and melt easily rationalizes trace element features such as the r.e.e. patterns, (garnet, in particular, is a probable cause of La/Yb fractionation), but variable ratios of Cs, Rb, K, Ba, U and Th to La are less obviously rationalized, even if the lattice sites provided by accessory minerals are involved. In particular, at least two accessory phases stable in equilibrium with the melt are required, because crystal sites for both U and Th, and the alkali metals and Ba are not found in common accessory phases (e.g. zircon has no site for alkali metals and phlogopite has no site for U and Th). Therefore, if alkali metals and U and Th are fractionated from rare earths, at least two accessory minerals are involved. Uranium-series studies indicate that Th/U ratios increase by up to 30% during melting of oceanic basalts, but no definitive mineralogical explanation for this fractionation has been proposed (see Allègre & Condomines 1982).

The trace element ratios of highly incompatible elements vary among o.i.b.: some of this variation is depicted in figure 1. The normalized levels of the elements Cs to Th are frequently less than the normalized value of La in the o.i.b.: a feature that they have in common with m.o.r.b. Note that the Gough Island sample is most similar to island arc basalts: it has the highest Cs and Rb to La ratios and is also isotopically extreme (see Kurz *et al.* 1982; Zindler *et al.* 1982; and figure 1).

PERIDOTITE SOURCE OF ISLAND ARC BASALTS

In oceanic island arcs, basalt derived from mantle peridotite is a common, and often predominant, magma type (see, for example, Kay *et al.* 1982). The basalts and andesites that erupt at arc volcanoes have often undergone crystal fractionation and mixing much as the m.o.r.b. and o.i.b. have. But, as with m.o.r.b. and o.i.b., as long as fractionation-independent chemical characteristics can be identified, inferences about the peridotite source composition can be made. Much the same can be said for various types of back-arc basin volcanism, probably including some 'flood basalts' (a term that is tectonically ill defined). For some convergent plate margin settings (especially Andean andesites) we may not be able to identify any surviving fractionation-independent characteristics (see, for example, DePaolo 1981; James 1982; Thorpe *et al.* 1981): the discussion in this section does not include these magmas.

Figure 2 illustrates some trace element characteristics that appear to be common in many island arc basalts. These include the higher ratios of elements grouped to the left (Cs, Rb, K, Ba, Th, U) to the largest r.e.e. (La). Armstrong first pointed this out (see Armstrong 1981 for a recent synthesis), and more recently Kay (1980) has also used Ba/La (see figure 3) and Pb/La ratios as evidence in favour of sediment subduction: a highly controversial postulate. The controversy involves an important question. Are the ratios of large cations (e.g. Cs/Ba, Ba/La) the same in basalts and in the mantle source (see Stern 1979)? This is of course the same question that was raised and skipped over (with almost unanimous consent) in the case of oceanic basalts. What must be claimed, if the trace element ratio differences between oceanic and island arc basalts are accepted, but if their peridotite sources have identical ratios, is that the mineralogically controlled fractionation differs in arc and oceanic magma generation. For the sake of discussion, assume that only minor large cation fractionation occurs in the oceanic case (m.o.r.b. or o.i.b.). Then, to explain the higher Ba/La ratios (for example) in arc basalts, by the process of partial melting one must hold that an accessory phase with high La distribution coefficient (and low Ba distribution coefficient) is stable. Kay (1980) has delineated the range of bulk distribution coefficients and melting percentages necessary to produce the desired increase in Ba/La ratio. Perovskite has been suggested as a high distribution coefficient residual phase



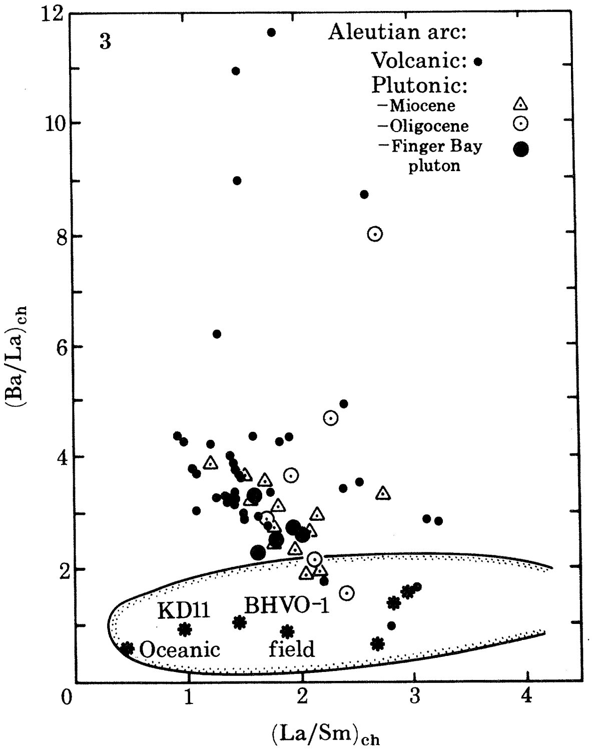


FIGURE 2. Normalized concentrations (see figure 1) of some trace elements in basalts from the Aleutian arc and the Tonga–Kermadec arc. A fundamental distinction between m.o.r.b., o.i.b. and i.a.b. cannot be made using the r.e.e., which show similar concentration levels and variability (compare figure 1 with 2). Normalized concentrations ratios of elements on the left side of the figure to La (the largest and least compatable r.e.e., when a Ca lattice site in minerals coexisting with melt is assumed) are almost invariably greater than one (solid pattern). Figure 1 shows that the same ratios are less than one (striped pattern) for most m.o.r.b. and o.i.b. The ratios are highest for the elements Ba and Cs, and lowest for Th. These differences between i.a.b., and m.o.r.b. and o.i.b. are probably due to differences in the concentrations in the source peridotites.

FIGURE 3. Concentration ratios Ba/La and La/Sm (chondrite-normalized) for the seven oceanic basalts (crossed Xs) plotted in figure 1, and a number of Aleutian arc volcanic and plutonic rocks (various symbols after Kay *et al.* 1983). Most of the island arc points fall outside of the oceanic field, at higher Ba/La values. Lowest Aleutian arc point is 'magnesian andesite' of Kay (1978), which is of doubtful tectonic classification.

(see, for example, Morris & Hart 1983). Generally, the higher U/Th ratios in arc basalts (compared with m.o.r.b. and o.i.b.) could be explained by distribution coefficients in perovskite; they are higher for Th than for U. But perovskite has a higher distribution coefficient for Th than for La (Onuma *et al.* 1981) and one would predict a very low Th content (and Th/La ratio) in the melts. This is not observed.

By way of positive argument one would like to find a peridotite with accessory phases that have high enough La concentrations to have low Ba/La ratios complimentary to the high

Ba/La in arc rocks. For the cases where mantle peridotites from arcs have been recovered (Japan and the Aleutians) this has not been observed (Tanaka & Aoki 1981; Kay, unpublished data, Aleutian Islands). Experimental studies of natural basalts with the K, Ti, P levels of arc basalts, has not substantiated the proposition that any of them could be at equilibrium with r.e.e.-bearing accessory minerals like apatite (Watson 1980; Watson & Green 1981) zircon, monazite or perovskite. In total, the foregoing observations do not support the postulate that large cation trace elements have been fractionated by accessory phases in island arc basalt genesis. The pressure–temperature stability of the accessory phases at mantle conditions is not in question: only their equilibrium with arc basalt.

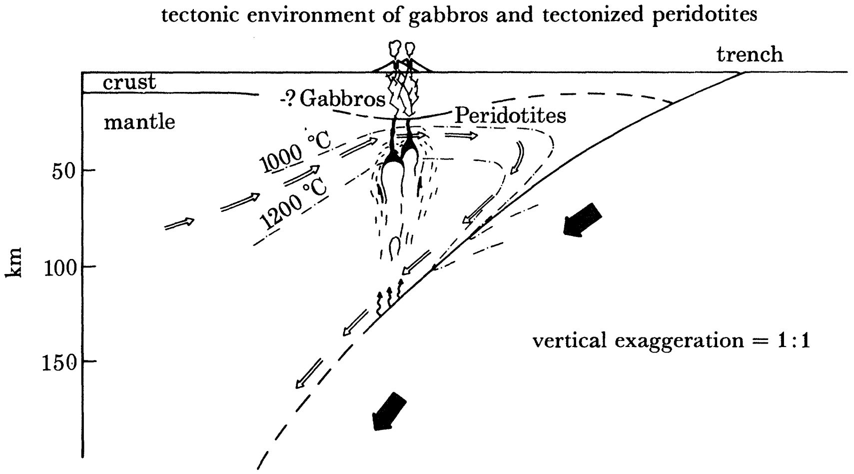
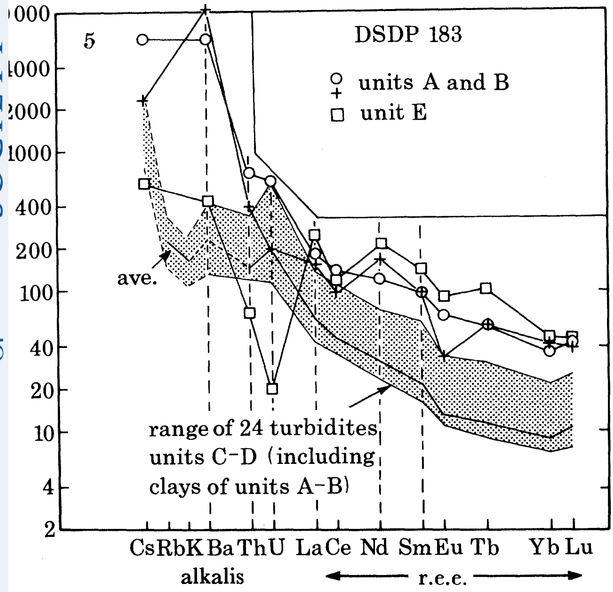


FIGURE 4. Geometrical arrangement of potential magma sources in an oceanic island arc (after Pope *et al.* 1983). Mantle xenoliths (tectonized peridotites) brought up by arc magmas constrain thermal structure. Isotopic analyses of xenoliths (Zashu *et al.* 1980; Kay & Wasserburg, in preparation) indicate that mantle with isotopic composition unlike that of the host magma overlies deeper source of diapirs.

The alternative to minor phase control of the differences in Ba/La ratio (for instance) between i.a.b., m.o.r.b. and o.i.b. is to postulate that the source peridotite itself is different and that this type of peridotite is uniquely associated with plate subduction. The juxtaposition of tectonically (and compositionally) diverse rocks that are possible components of arc magmatism is shown in figure 4. In a general way, it has long been recognized that oceanic sediments have some of the required trace element characteristics to serve as an end member for arc volcanism (see Armstrong 1971; Dupuy *et al.* 1982; Hawkesworth & Powell 1980; Hickey & Frey 1982; Karig & Kay 1981; Kay 1980; Perfit *et al.* 1980; Whitford *et al.* 1977). This is also true of the isotopic systems: (see, for example, DePaolo & Johnson 1979; Sun 1980). But the very diversity of the sedimentary column and the paucity of analyses of the select group of elements that reveal the i.a.b.-m.o.r.b. and o.i.b. differences, has made the trace element calculations used to substantiate sediment as a component in arc magmas very generalized and unsatisfactory.

New trace element data plotted in figures 5 and 6 (see table 3 for representative analyses) for North and Central Pacific sediment types, illustrate the diversity in sediment components of possible relevance to the Aleutian arc magma geochemistry. The samples in figure 5 are from a Deep Sea Drilling Project core that penetrates the entire sediment section at a locality immediately south of the Aleutian trench. Among the noteworthy geochemical features of the sediments are the extremely similar relative contents of the elements (e.g. the patterns are the same) in 24 Eocene continent-derived turbidites from the middle of the core. This trace element pattern has a high ratio of Ba/La and especially Cs/La. The diatomaceous sediments at the top of the core have even higher Ba/La and Cs/La ratios, and their Ba/Cs ratios are more similar to those of arc basalts than the Ba/Cs ratios of the turbidites are. Pelagic carbonate at



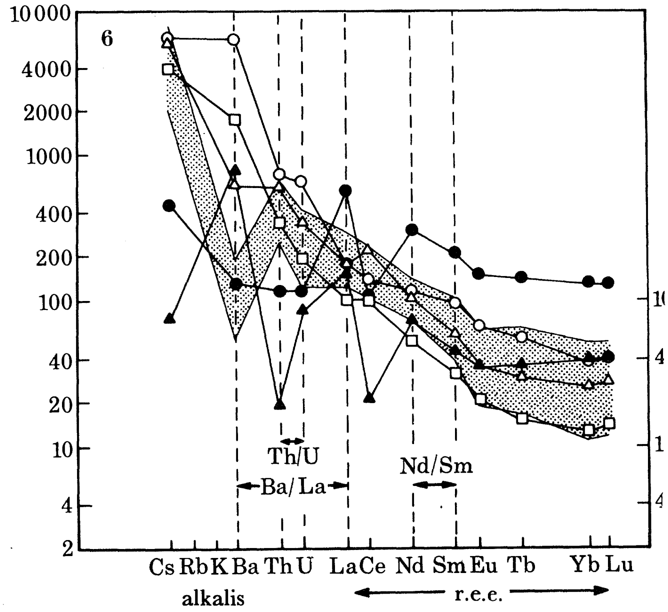


Figure 5. Normalized concentrations (see figure 1) of some trace elements in sediments from Deep Sea Drilling Project hole 183, (161° W, 53° N) located 100 km south of the Aleutian trench (see Stewart 1977). Units A and B, from the core top to Miocene are diatomaceous sediments with ash and glacial erratics. The Oligocene–Eocene age units C and D form a monotonous turbidite sequence. The core bottoms in Eocene alkali basalt (r.e.e. analysis in Kay 1980) with unit E, a pelagic carbonate overlying it. Eocene granites from the Shumagin Islands, 50 km north of hole 183 in the Aleutian forearc, have patterns generally within the field of turbidite sequence (Hill *et al.* 1981, see text for discussion). Representative analyses are in table 3.

Figure 6. Normalized concentrations (see figure 1) of some trace elements in various pelagic sediment types of the north and central Pacific ocean. Four zeolitic clays from the latitude of Hawaii (10°–30° N) shown by stippling have similar trace element patterns and are enclosed in a field, more northerly sediments (35°–55° N) shown by open symbols are diatomaceous and follow the Ba-enriched pattern illustrated at the top of DSDP hole 183 (figure 4). Slowly deposited equatorial pelagic sediments (10° S) shown by black symbols have negative Ce anomalies and high r.e.e. contents that are similar to that of the overlying water column. Representative analyses are in table 3.

the base of the core has a distinct negative Ce anomaly and low Th content, typical of metalliferous sediment overlying basaltic oceanic crust.

The core was taken on a seamount, thus the basal section of the core is probably not representative of pelagic sediment overlying m.o.r.b. crust (and this is the relevant basal sediment for much of the crust undergoing subduction at the Aleutian trench). Analysis of several pelagic sediment types from the central and northern Pacific are plotted in figure 6 (see table 3). Noteworthy features of the trace element patterns are: (i) the low Ba/La ratios and high Cs/Ba ratios of the zeolitic clays. These ratios are quite unlike those of arc volcanics. (ii) The negative

Ce anomalies and variable Ba and Cs contents (and Ba and Cs to La ratios of some pelagic clays and (iii) the high Ba/La ratios of radiolarian oozes. Generally, the sediment component that most resembles that postulated by Kay (1980) to represent the sediment end-member of Aleutian arc volcanism is a siliceous (radiolarian) pelagic clay from near to top of the sediment column. This is of some concern, for one would expect basal sediments to be preferentially subducted, and they do not appear to have the right trace element ratios (e.g. Cs/Ba) to serve as components of arc magmas. The recent discovery of ^{10}Be in arc volcanics was similarly unexpected for it implies subduction of some of the younger parts of the sediment section (see Brown *et al.* 1982). Perhaps the trace element patterns and ^{10}Be are telling us the same thing.

Table 3. Trace elements in terrigenous and pelagic deep-sea sediments

	1	2	3	4	5	6	7	8
Ba	919	2437	781	169.8	8770	442	892	603
La	14.2	7.04	24.8	52.3	174	40.5	22.1	10.8
Ce	30.5	14.5	51.3	72.6	62.2	84.8	42.7	29
Nd	21.2	7.53	20.0	55.4	154	48.1	18.1	14
Sm	4.55	2.36	5.32	14.4	30.3	7.35	4.63	3.87
Eu	1.20	0.592	1.31	2.51	8.6	1.38	1.30	0.183
Tb	0.685	0.346	0.751	1.78	7.9	0.11	0.78	0.76
Yb	2.60	0.972	3.91	5.04	29.2	2.37	2.47	3.45
Lu	0.417	0.166	0.401	0.655	4.62	0.400	0.258	0.505
Cs	2.54	1.00	4.02	6.67	1.00	8.4	5.2	9.0
Th	3.42	1.72	8.22	11.1	2.72	11.7	5.83	6.7
U	2.51	0.885	2.74	1.50	4.02	2.64	2.54	3.26
Hf	3.15	1.70	4.26	3.80	3.62	3.51	3.84	2.8
Ta	0.469	0.136	0.78	1.20	0.218	0.99	0.734	0.70
FeO	6.77	1.42	6.60	5.8	22.1	6.03	6.20	1.28

(1) DSDP 183/2/3/50–52 diatomaceous silty clay.
(2) DSDP 183/17/3/70–72 diatomaceous ooze.
(3) DSDP 183/28/2/133–153 turbidite sequence.
(4) DSDP 170/7/3/5/3–55 zeolitic brown clay.
(5) DSDP 319/1/6/130-132 Fe-rich brown clay.
(6) MAG-1 Marine mud USGS standard.
(7) S89d Argillite, Shumagin Formation Sanak Island, Hill *et al.* (1981).
(8) N170 Granophyric Granite Shumagin Batholith, Hill *et al.* (1981).

Saunders *et al.* (1980) and Thompson *et al.* (1984) note the distinctively low ratios of Ta and Nb to La in volcanic arc magmas (relative to m.o.r.b. and o.i.b.). Ta and La concentrations of Aleutian island volcanic rocks (table 2 and my unpublished data) show the same geochemical signature. Saunders *et al.* (1980) attribute the low ratios to retention of Ta and Nb in titaniferous mineral phases such as ilmenite and sphene in the actively dehydrating subducted oceanic crust (transformed to eclogite). Thompson *et al.* (1984) concur with the explanation of Saunders *et al.* (1980), but point out that Ti does not always correlate with Ta and Nb, indicating that the residual phase may not be titaniferous. An alternative explanation is suggested by the low Ta to La ratios found in the oceanic sediments listed in table 4. Following Kay (1980), one may estimate that about one half of the La in the peridotite source of Aleutian arc magmas is derived from subducted sediment. This conclusion is based on the Nd isotope composition of arc magmas, which is interpreted as a mixture of m.o.r.b. and sediment components. In contrast to La, the sediment will contribute 30 % or less of the Ta in the mixture, which will tend to have a low Ta to La ratio, and therefore could serve as a source for arc magmas.

General discussion of peridotite sources

The identification of mantle heterogeneity in both isotope and trace element systematics raises the question of the causes of the heterogeneities. The subarc mantle appears to be an ideal locality to create mantle heterogeneity. As shown in figure 4, the entrainment of adjacent asthenospheric mantle by the subducting lithospheric plate brings a continuous flow of mantle peridotite through the plane of magma production, shown as diapirism starting at a thermally and chemically perturbed region near the subducting plate. Over Earth history a significant proportion of upper mantle peridotite has been through arc processing of the sort depicted in figure 4. For instance, at a flow rate of only 2 cm a^{-1} a thickness of 30 km of mantle processed at the 40000 km length of subducting plates results in 2.4×10^{16} cm^3 of mantle processed each year. At this rate about one fifth of the upper mantle volume (2.4×10^{26} cm^3) could have been through processing at arc regions over the past 2 billion years. Hofmann & White (1982) have proposed that subduction is important in creating the long-lived heterogeneities that (following billion-year-scale ageing in the mantle) become source regions for intrapolate volcanism (the o.i.b. of figure 1). The end member represented by Gough Island in figure 1 has some of the trace element characteristics and isotopic characteristics of arc volcanic rocks, but other o.i.b. types seem to represent distinct geochemical groups (as argued above) and for that reason I am reluctant to accept the close similarity of most o.i.b. and i.a.b. source peridotites even allowing for isotopic ageing (which of course does not affect the trace element ratios). But the general argument of Armstrong (1981) that the isotopic and trace element development of the upper mantle requires continuous mass input at island arcs is in agreement with the sediment inputs to arc magmatism inferred here, perhaps in the style suggested by McKenzie & O'Nions (1983).

Crustal sources: three types

Melting of oceanic crust is a plausible mechanism for production of some magma types in oceanic island arcs. The geometry is that depicted in figure 4, but now the requirement of equilibrium with olivine (peridotite) is removed. Unless very large percentages of melting are encountered, which is unlikely due to probable segregation of the melt, the composition of the melt will not be basalt, let alone basalt with substantial Ni and Cr content, such as that found in several island arcs (see, for example, Kay *et al.* 1982; Conrad & Kay 1984). There are andesites in island arcs that appear to be primary melts (e.g. they are not derived from or even closely associated with Ni- and Cr-rich basalts: Kay 1978; Aoki & Fukimaki 1982; Tatsumi & Ishizaka 1981, 1982). The trace elements and isotopes in some of these andesites have been plausibly modelled as small-percentage partial melts of subducted hydrothermally altered (Hart & Staudigel 1982) oceanic crust (see Kay 1978), although the alternative hydrous melting of the overlying peridotite now seems more likely (Tatsumi & Ishizaka 1981, 1982).

The melting of continental crustal material is plausible wherever basaltic magma enters the continental crust (see review of Kay & Kay 1981). The compositional diversity of the products of crustal melting is reduced by the requirement of a eutectic-type (granitoid) composition. Thus, as Barth (1968) pointed out, melting simplifies the composition rather than differentiating it. The discovery of abundant metamorphosed basic igneous rocks in the lower crust (sampled as xenoliths) of island arcs, makes remelting of lower crustal igneous rocks (in a differentiation-assimilation process: DePaolo 1981) a likely process resulting in the I (for igneous)-type granitoids commonly found at convergent plate margins. For oceanic island arcs, the young crustal age makes this remelting difficult to detect isotopically, and the trace element patterns of the melt may resemble those of closed system fractionation of basalt at the same depth. For instance, the granitoids (largely diorite and granodiorite) of the Aleutian islands share the same geochemical features as mantle-derived basalts (see figure 3). Only when crustal ageing creates isotopically distinct crustal regions will the melting be identified (see, for example, Hawkesworth *et al.* 1983).

One type of crustal source appears not to have yielded any magmas by partial melting: the residual crust represented by high grade granulites (see, for example, Sheraton *et al.* 1973). The distinctive features of the granulites crust are the low U and Th content (see table 1) and Rb/Sr ratios for rocks of intermediate to basic composition. These characteristics may have been caused by removal of an igneous melt or fluid phase. No magmas have the trace element characteristics of the granulites (even allowing for fractionation during melting), although the lower crustal granulites may be components in intermediate composition magmas that have undergone assimilation-crystal fractionation in the crust. One might identify this R-type (for residual) granitoid component by its distinctively low U, Th and Rb to r.e.e. ratios. On the other hand, the feldspar in the R-type sources provides a site for Sr and Ba, and Ba/La ratios are not low in these rocks (see table 1).

A third type of crustal source is sediment. This has already been discussed as a component in mantle peridotite at subduction zones, but now the discussion will focus on a series of 60 Ma old silicic S-type granitoids that intrude accretionary prism metasediments in the Aleutian forearc (Hill *et al.* 1981). Table 3 shows that the distinctive trace element chemistry of these forearc plutons: mutual ratios of Cs, Ba, U and Th agree with the ratios in argillite that is cut by the plutons (also listed in table 3) and with the 'turbidite sequence' sediments of somewhat younger age (Eocene) in DSDP hole 183, south of the Aleutian trench (see above). Sr and O isotope studies of Hill *et al.* (1981) reveal that a basaltic component (lower $^{87}Sr/^{86}Sr$, lower $\delta^{18}O$) is required in addition to the sediment component, but the geochemical signature of the alkali metals, Ba, U, Th and the light r.e.e. is clearly dominated by the sediment.

The compositional (and isotopic) diversity of continental crustal magma sources lends a particularly *ad hoc* flavour to the discussions that are based purely on element systematics in the melt products themselves. Theoretical calculations based on trace element distribution coefficients assume equilibrium and are critically dependent on the melting temperature and dissolved volatiles. Fractionation of minor minerals (e.g. monazite) may remove elements (e.g. l.r.e.e.) that are regarded as incompatible in basaltic systems (see Mittlefehldt & Miller 1983; Michael 1983). The residual mineralogy of the 'restite' has been parametrized almost at will by some workers to fit geochemical models. Recovery of fragments of lower crustal material among the more frequently studied mantle xenoliths, opens the questions of temperature, mineralogy and disequilibrium to direct study.

SUMMARY

Trace element heterogeneity in the source regions of magmas as diverse as basalts and leucogranites can be established using analyses of fractionation-independent elements of the magmas. In general, crystal-melt fractionation is not an adequate explanation for much of the variability in the sources. Mantle metasomation (Bailey 1982) or analogous crustal metasomatism (Heier 1979), involving a high pressure CO_2- or H_2O-rich fluid phase may create

some of the heterogeneity, but high concentration levels of many elements in the fluid are required.

The recycling of sediment with distinctive and highly variable ratios of Cs, Rb, K, Ba, U, Th and r.e.e., which recent ^{10}Be studies (Brown *et al.* 1982) have reaffirmed, creates new opportunities for quantitative mass balance calculations on crustal recycling. More emphasis on sediment chemistry (including regional variability) and on the chemistry of mantle and crustal residues (restites) will be necessary to further the at present rather generalized and speculative aspects of the recycling calculation.

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