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Chapter 22

Aleutian magmas in space and time

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INTRODUCTION

The Aleutian arc provided the setting for a proposal by Coats (1962) that arc magmas are related to a subducted oceanic plate in a convergent tectonic setting—a proposal that predated the theory of plate tectonics. Since that time, the study of Aleutian arc magmas has generated controversies over their origin and evolution that are fundamental to understanding the origin of magmas in all island arcs.

The first and longest controversy has been over the composition of the most important parental lava (precursor to most arc magmas) derived from the mantle. The question is whether this magma is a high-Al basalt generated by partial melting of the subducting plate (e.g., Marsh, 1982) or a more Mg-rich basalt derived from the peridotite overlying the plate that has been fluxed by a component from the plate (Kay, 1977; Perfit and others, 1980b).

The second major controversy is over the origin of the arc-type trace-element and isotopic characteristics of Aleutian magmas. The models considered for the magmatic source include (1) a mixture of subducted oceanic crustal and sedimentary components with a depleted mid-oceanic ridge-type mantle (Kay and others, 1978; Kay, 1980); (2) a subducted oceanic crustal source (Marsh, 1976) that includes a subducted sedimentary component (Brophy and Marsh, 1986); and (3) an upper mantle source composed of mid-oceanic ridge and oceanic island-type components (Morris and Hart, 1983).

The third major controversy is over the origin of the petrologic and geochemical diversity along the arc and within individual centers, and the relation of this diversity to differences in tectonic setting along the arc. The petrologic questions here involve the depth of fractionation in the crust, the role of magma mixing, the role of water, and the importance of crustal melting, assimilation, and incorporation of xenolithic fragments of the preexisting crust and mantle.

The aim of this chapter is to review the petrology and geochemistry of Aleutian arc magmas in order to present an overview of the regional variations in the arc and to identify the processes responsible for the origin and evolution of the magmatic rocks that constitute the arc crust. This overview leads us to a coherent model for the origin of the Aleutian crust. The petrologic and geochemical changes through time in the central part of the arc will also be briefly described.

DISTRIBUTION AND GENERAL CHARACTERISTICS OF ALEUTIAN ARC MAGMATISM

The geologic framework of the Aleutian arc, with emphasis on the offshore regions, is discussed by Vallier and others (this volume), and the general distribution and physical characteristics of Aleutian volcanoes have been summarized by Wood and Kienle (1990), Fournelle and others (this volume), and Kay and others (1982). Consequently, these aspects of the arc will not be discussed here. Instead, the purpose of this section is to review the aspects of the tectonic framework and physical characteristics of the volcanoes that are needed to understand the discussion of the petrology and geochemistry that follows.

As discussed by Marsh (1979), the main front of the Aleutian arc volcanoes can be grouped into linear segments based on geographic distribution. In the oceanic and western continental part of the arc, the segmentation scheme of Marsh (1979) was modified by Kay and others (1982) who defined the Rat, Andreanof, Four Mountains, and Cold Bay segments (Fig. 1). The ends of these linear segments correlate with major tectonic breaks in the overlying and underthrust plates and coincide with the terminations of rupture zones of major earthquakes (Kay and others, 1982). From west to east, the boundary between the Rat and the Andreanof segments coincides with the intersection of the Bowers ridge with the arc, the boundary between the Andreanof and Four Mountains segments coincides with the Amlia fracture zone on the lower plate, and the transition zone between the Four Mountains and Cold Bay segments coincides with the transition

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from oceanic to continental crust. See Vallier and others (this volume) for further discussion of these features, and Kienle and Swanson (1983) for discussion of segmentation in the eastern Aleutian arc.

The region from the central part of the Cold Bay segment through the Four Mountains segment (Fig. 1) has several other complexities. First, for some 200 km between the Four Mountains and Cold Bay segments, the volcanoes do not form coherent lines. Second, volcanoes of the Four Mountains segment are closer to the trench than those of other segments. Finally, a secondary volcanic front consisting of small centers (such as Bogoslof, Amak, Uliaga, Kagamil, Carlisle, and smaller unnamed submarine centers) is developed behind the main arc in this region.

Variations in size, morphology, and petrologic characteristics of Aleutian arc volcanoes are related to position within the arc. As discussed by Marsh (1979), volcanic centers in the eastern part of the arc tend to be larger than those in the western part.

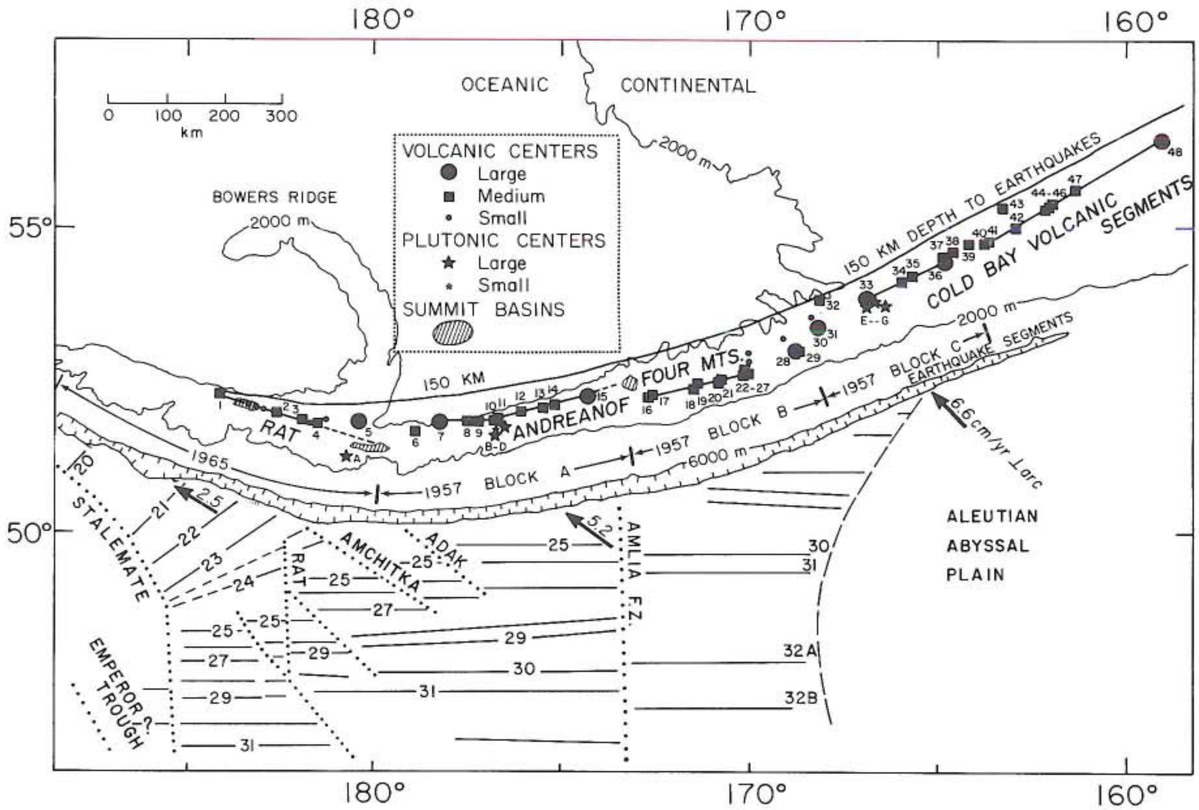


Figure 1. Bathymetric, tectonic, and magmatic features of the volcanically active part of the Aleutian Island arc. The figure is modified from Figure 1 of Kay and others (1982). Modifications include changes in the magnetic anomaly pattern (lines with numbers) on the Pacific Plate (from Lonsdale, 1988) and in the perpendicular convergence rates (from Engebretson and others, 1985). Volcanic (numbered) and plutonic centers (lettered) are listed below and are labeled as calc-alkaline (CA), tholeiitic (TH), or unknown or ambiguous (?) based on whole-rock composition, and mineralogy. Numbers in parentheses indicate number of whole-rock analyses available. Volcanoes: 1, Buldir (CA) (8); 2, Kiska (CA) (4); 3, Segula (TH?) 2; 4, Little Sitkin (CA) (13); 5, Semisopochni (TH) (15); 6, Gareloi (?); 7, Tanaga (?); 8, Bobrof (CA) (7); 9, Kanaga (CA) (20); 10, Moffett (Adak I.) (CA) (>20); 11, Adagdak (Adak I.) (CA) (10); 12, Great Sitkin (CA) (26); 13, Kasatochi (CA) (3); 14, Koniuji (CA); 15, Atka (TH) (>20); 16, Unnamed (Seguam I.) (?); 17, Unnamed (Seguam I.) (?); 18, Amutka (CA) (1); 19, Chagulak (TH?) (1); 20, Yunalaska (TH) (1); 21, Unnamed (Yunalaska I.); 22, Herbert (TH) (1); 23, Carlisle (CA) (1); 24, Cleveland; 25, Chuginadak (?) (1); 26, Kagamil (?) (1); 27, Uliaga (CA) (1); 28, Vsividof (Umnak I.) (TH) (4); 29, Recheschnoi (Umnak I.) (?) (4); 30, Okmok (Umnak I.) (TH) (>20); 31, Tulik (Umnak I.) (TH); 32, Bogoslof (CA) (3); 33, Makushin (Unalaska I.) (TH-CA) (14); 34, Akutan (TH) (10); 35, Akun (?); 36, Faris-Westdahl (Unimak I.) (TH) (7); 37, Pogromni (?); 38, Fisher (?); 39, Shishaldin (TH); 40, Isanotski (?); 41, Unnamed (?); 42, Frosty and Mt. Simeon (CA) (>10); 43, Amak (CA) (7); 44, Emmons (?); 45, Pavlof (TH): 46, Pavlof Sister (?); 47, Dana (?); 48, Veniaminof (TH) (4). Volcanoes to the east, including Augustine, are classified by Kienle and Swanson (1983). Plutons: A, East Cape (Amchitka I.) (3) (CA); B, Finger Bay (Adak I.) (TH) (26); C, Hidden Bay (Adak I.) (CA) (22); E, Shaler (Unalaska I.) (CA) (7); F, Captains Bay (Unalaska I.) (CA) (29); G, Beaver Inlet (Sedanka I.) (CA) (8).

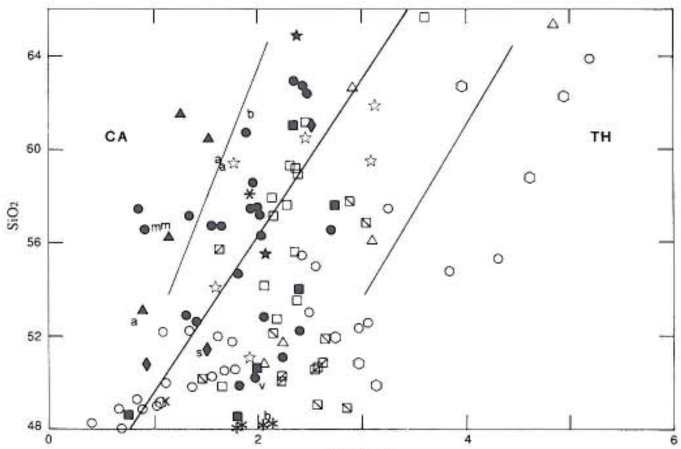


Figure 2. Plot of SiO_2 versus FeO* (total Fe as FeO)/MgO ratio for Pliocene-Recent Aleutian arc volcanic rocks for which trace-element data exist. A heavy line (from Miyashiro, 1974) separates the calc-alkaline (CA) from the tholeiitic (TH) field. Other lines separate the calc-alkaline from the transitional calc-alkaline field and the transitional tholeiitic from the tholeiitic field (following S. Kay and Kay, 1985a). Symbols: solid triangles, Buldir; solid stars, Little Sitkin; s, Segula; asterisks, Kanaga; solid circles, Moffett (m - high-Mg andesites of Kay, 1978); solid squares, Adagdak; K, Kasatochi; solid diamonds, Makushin; open squares, Great Sitkin; open stars, Islands of Four Mountains; open squares (NE-SW slash), Atka; open squares (SE-NW slash), Akutan; open triangles, Semisopochnoi; v, Veniaminof; open circles, Okmok; open hexagons, Westdahl; b, Bogoslof; a, Augustine. Data from Kay (1977), Kay and others (1982), Arculus and others (1977), Byers (1961), McCulloch and Perfit (1981), Delong and others (1985), Myers and others (1986b), Neuweld (1987), Nye and Reid (1986), and Appendix (see microfiche accompanying this volume).

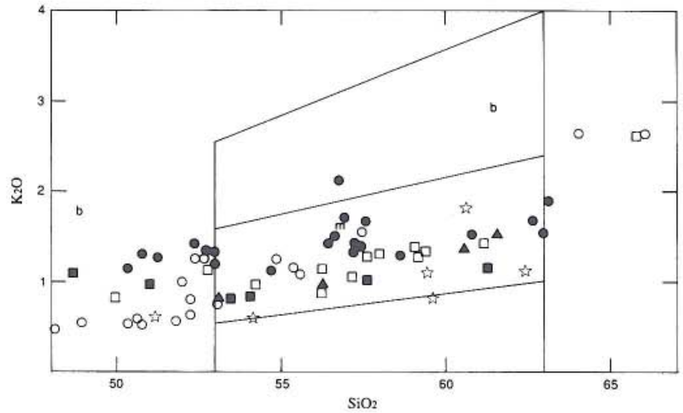


Figure 3. Plot of K_2O versus SiO_2 for selected Aleutian volcanic centers. Low, medium-, and high-K andesite fields of Gill (1981) are delineated. Symbols are the same as in Figure 2. Main arc tholeiitic andesites, and especially dacites, tend to have higher K_2O (Okmok, open circles; Westdahl, open squares) than main arc calc-alkaline andesites and dacites (solid circles, Moffett; solid triangles, Buldir; open squares, Great Sitkin; solid squares, Adagdak). Four Mountains samples (stars), except Uliaga, have low K; Bogoslof samples (b) have high K.

These size differences correlate with, and probably are due to, the decreasing perpendicular convergence rates to the west resulting from the oblique convergence of the Pacific Plate (Fig. 1). Superimposed on these east to west variations are size variations related to the location of volcanic centers within the segments. In general, volcanic centers at the ends of segments and between segments are larger than centers in the middle of segments. These variations are particularly well developed in the central and western part of the arc (Fig. 1). Compositionally, the largest volcanoes are predominantly basaltic with lesser amounts of nearly aphyric andesite and dacite; the smallest volcanoes have fewer basalts and much higher percentages of porphyritic andesite and dacite.

Chemical trends differ in volcanic rock series from large and small volcanic centers. By the criteria of Miyashiro (1974), lavas from the large centers are largely tholeiitic (that is, their FeO/MgO ratios increase with SiO_2), whereas those in the small centers are calc-alkaline (their FeO/MgO ratios are relatively constant with increasing SiO_2; Fig. 2). Based on this classification, volcanic rocks of the arc form a compositional continuum between tholeiitic and calc-alkaline characteristics. Following S. Kay and Kay (1985a), samples within this continuum will be designated tholeiitic, transitional tholeiitic, transitional calc-alkaline, and calc-alkaline (Fig. 2). The centers used as references will be: Buldir, a calc-alkaline center; Okmok and Westdahl, tholeiitic centers; and Moffett, Great Sitkin, and Semisopochnoi (also Atka), a sequence of centers that progress from transitional calc-alkaline through transitional tholeiitic centers. As shown in Figure 2, lavas within a single center can fall within several of these classifications (e.g., Moffett), but most often lavas above 54 percent SiO_2 fall within one group. The aim of this classification is to draw attention to the diversity of magmatic characteristics that require a diversity of processes.

Chemical differences in Aleutian arc magmas are also observed perpendicular to the arc. As in many arcs, these differences are shown by an increase in K_2O content at a given SiO_2 content with distance from the trench (see Figs. 1 and 3). When comparing the K_2O content of several samples, Kay and others (1982) showed that the Mg number (molar ratio of Mg to Mg + Fe) has to be considered, since Aleutian tholeiitic andesites have a higher K_2O content than calc-alkaline andesites with the same SiO_2, because of different fractionating mineralogies (see Fig. 3). For example, Bogoslof volcano, which is located north of Okmok volcano and is the largest of the centers behind the arc and the farthest from the trench, has the highest K_2O content of any center in the arc (Byers, 1961; Arculus and others, 1977). Similarly, lavas from Amak, north of the Cold Bay volcanic center, have a higher K_2O content than Cold Bay lavas (Marsh and Leitz, 1979). In contrast, Four Mountains volcanic centers, which are the closest to the trench (and presumably the closest to the seismic zone), have lavas with the lowest K_2O content. Four Mountain segment centers located behind the main volcanic front have higher K_2O content. The widening of the area of arc volcanism and the chemical diversity in this region are yet to be explained in any comprehensive Aleutian magmatic model.

Although the K_2O content of volcanoes decreases trenchward across the Aleutian volcanic arc, no low-K island arc tholeiites, such as those occurring in the frontal arc of the western Pacific island arcs (Gill, 1981), have been found in the Aleutian arc. One other important, and possibly related difference between the western Pacific arcs and the Aleutians is that the back-arc region in the Aleutians is a trapped piece of Mesozoic oceanic crust (Vallier and others, this volume) rather than a region where back-arc spreading is occurring or has occurred.

MODELS FOR THE ORIGIN OF MAGMATIC DIVERSITY IN THE ALEUTIAN ARC

Two basic models have been presented to explain the magmatic diversity in Aleutian volcanic rocks. In one model, Kay and others (1982) and S. Kay and Kay (1985a, b) have argued that the most common parental magmas are olivine tholeiitic (Mg-rich) basalts derived from mantle peridotite that has been modified by material from the subducted plate. In this model, a common process in all Aleutian volcanoes is the formation of high-Al basalt from parental olivine tholeiite at the base of the crust. Major-element and compatible trace-element variations and mineralogic differences between Aleutian magma series then result from fractionation and mixing processes in the Aleutian crust that are controlled by the tectonic setting of the volcano. Magmas that feed the high-volume tholeiitic centers pass quickly through crust that is under relative extension, resulting in shallow-level magma chambers where crystallization occurs at low pressure and there is little mixing or interaction with the arc crust. In contrast, magmas that feed the lower volume calc-alkaline centers pass slowly through crust under relative compression, and undergo crystallization and mixing at greater crustal depths. Important components of this model are correlation of volcanic volume with center type, and the observation that the smaller calc-alkaline centers are located in the middle of tectonically controlled arc segments, whereas the larger tholeiitic centers are located at the ends of these segments or in regions where segments are not well defined. Exceptions occur in the Cold Bay segment, where transitional tholeiitic lavas are common in intrasegment volcanoes such as at Akutan (Romick, 1982). However, in this region the largest volcanoes with the most extreme tholeiitic characteristics still occur at the end of or between segments.

In the second model, developed by Marsh and his coworkers (e.g., Myers and others, 1985, 1986b), differences along the arc are related to the relative age of the centers. In this model, melting of the down-going oceanic crust plus subducted sediment yields parental high-Al basalts, and crustal differentiation of high-Al basalts is a factor in creating the compositional spectrum observed in the volcanic centers. Calc-alkaline centers ("dirty" centers) form in an early immature stage of conduit development when significant lithospheric mantle debris is incorporated in the magma during transit through the lithosphere. In time, the volcanic center transforms into a tholeiitic center ("clean" center) as the conduit is thermally and chemically preconditioned and no longer contaminates the magmas. Duration of activity at a center explains the size differences between the tholeiitic and calc-alkaline centers. A problem with this model is that its evolutionary sequence is based on a comparison of volcanoes along the arc, since no volcano that has been studied shows a progression from calc-alkaline to tholeiitic lavas. On the contrary, the earliest (2 Ma) lavas from the large tholeiitic Okmok volcano (Bingham and Stone, 1972) are not calc-alkaline (Byers, 1961; Kay and others, 1982).

Clear distinctions between these two models involve different views of the evolution of Aleutian magmas in the arc crust and mantle, and different views of mantle dynamics beneath the arc. Differences in views on mantle dynamics are reflected in the relation between the arc lithosphere and asthenosphere in the two models. In the second (Marsh) model, the lithosphere includes much of the mantle above the down-going slab as well as the arc crust. In the first (Kay) model, most of the mantle above the down-going slab is undergoing ductile flow and is referred to as asthenosphere. DeBari and others (1987) suggest that this asthenospheric flow extends almost to the base of the crust, because of the penetratively deformed cumulate dunitic and wehrlitic xenoliths from Adagdak volcano (Adak Island) that are thought to have formed just below the crust.

Origin of Aleutian primitive and parental magmas

Fundamental to resolving the origin of the diversity in Aleutian arc volcanic rocks and the origin of island-arc crust, is an understanding of the formation of primitive and parental arc magmas. In this context, primitive basalts are those whose compositions are in equilibrium with the upper mantle. Primitive basalts lack evidence of crustal mixing, and those that have been at equilibrium with mantle peridotite have high Mg numbers and Cr and Ni content, indicating no (or more realistically, minimal) crustal-level fractionation. Mg-rich basalts (>8.5 percent MgO) that satisfy these criteria erupt infrequently in the Aleutians, but occur in both the calc-alkaline and tholeiitic centers. In contrast, important parental magmas are those magmas that evolve to produce the majority of the arc lavas. In this context, the most abundant basalts in the Aleutians, high-Al basalts (17 to 21 percent Al_2O_3), are important parental lavas in both models outlined above. Representative analyses of both Mg-rich and high-Al basalts are given in Tables 1 and 2 and plotted in Figure 4. The interrelationship of the major-element compositions of these basalts is critical to understanding the origin of the high-Al basalt type. Variations in incompatible trace-element ratios of both Mg- and Al-rich basalts reflect spatial and temporal differences in the magmatic source region that are largely decoupled from variations of major-element composition.

Mg-rich basalts (see Table 1) have been described from the tholeiitic center of Okmok (Byers, 1959, 1961; Kay and others, 1982; Nye and Reid, 1986), the calc-alkaline centers of Makushin (Perfit and others, 1980a), Kasatochi (S. Kay and Kay, 1985a) and Adagdak (DeBari and others, 1987), and from Ter-

TABLE 1. REPRESENTATIVE ANALYSES OF Mg-RICH AND HIGH-Mg ANDESITE FROM THE ALEUTIAN ISLANDS, ALASKA

Sample	OK1	OK1A	OK4	MK15	ADH	KAN2	KAS7A	MOF53A
SiO_2	45.70	48.07	48.97	51.20	48.46	50.28	49.07	55.50
TiO_2	0.51	0.73	0.72	0.75	0.69	0.75	0.66	0.86
Al_2O_3	12.90	15.64	16.27	15.69	15.14	16.10	15.75	15.50
Fe_2O_3	3.60	2.34	2.83					
FeO	5.60	6.98	6.24	9.21	9.03	8.74	8.82	6.21
MnO	0.03	0.15	0.01	0.18	0.14	0.13	0.14	0.10
MgO	18.20	12.68	9.62	9.64	11.83	11.19	8.77	5.58
CaO	9.60	10.76	12.86	10.12	11.37	9.99	12.52	9.51
Na_2O	1.83	2.14	2.11	2.77	2.12	2.57	2.24	2.98
K_2O	0.52	0.48	0.54	0.92	0.74	0.58	0.76	1.47
P_2O_5	0.10	0.12	0.13	0.21			0.15	0.32
Total	98.59	100.09	100.30	100.69	99.52	100.33	98.88	98.03
La	3.93	7.30	4.29	7.32	5.85	5.66	5.43	29.15
Ce	9.45	18.15	11.03	18.30	13.28	14.91	13.61	64.11
Nd	6.41	11.38	6.40	12.80	9.03	9.05	7.74	34.06
Sm	1.73	2.69	2.00	3.04	2.21	2.55	2.34	5.77
Eu	0.577	0.834	0.696	0.870	0.720	0.841	0.652	1.638
Tb	0.324	0.447	0.374	0.460	0.370	0.477	0.420	0.495
Yb	1.23	1.52	1.29	1.55	1.31	1.78	1.49	0.94
Lu	0.195	0.223	0.199		0.207	0.257	0.219	0.137
Rb				11			12	
Sr				533			482	
Ba	132	241	164	265	222	207	228	536
Cs	0.37	0.06	0.32	0.82	1.05	0.25	0.56	0.24
U	0.60	0.70	0.45		1.08	0.43	0.65	1.29
Th	0.66	1.42	0.92		1.99	0.84	1.60	3.00
Hf	1.10	1.48	1.04		2.09	1.75	1.30	4.03
Ta	0.05	0.09	0.08		0.08	0.13	0.10	
Sc	41.3	37.9	49.4		44.5	40.3	44.7	20.9
Cr	1230	698	608	255	614	946	338	377
Ni	360	272	98	66	219	222	61	153
Co	60	51	45		41	43	41	
FeO/MgO	0.49	0.72	0.91	0.96	0.76	0.78	1.01	1.11
Ba/La	33.5	33.1	38.1	36.2	38.0	36.6	42.0	18.4
La/Sm	2.3	2.7	2.1	2.4	2.7	2.2	2.3	5.1
La/Yb	3.2	4.8	3.3	4.7	4.5	3.2	3.7	30.9
Eu/Eu*	0.98	0.95	1.03	0.91	1.00	0.97	0.83	1.09
Th/Hf	0.60	0.96	0.88		0.95	0.48	1.24	0.74
Ba/Hf	119	163	158		106	119	176	133
Ba/Th	200	170	179		112	246	143	179
La/Hf	3.56	4.93	4.13		2.79	3.24	4.19	7.23
Th/U	1.1	2.0	2.0		1.8	1.9	2.5	2.3
$^{87}Sr/^{86}Sr$		0.70302		0.70333				0.7028
ε_{Nd}		7.54		6.17				11.3
$\delta^{18}O$	4.9	5.8	5.2		5.2	7.9		7.5

*OK1, OK1A, and OK4 are from the flank of Okmok volcano, Umnak Island (Byers, 1961). Major-element analyses are mostly from Byers (1961) and isotopic analyses are from McCulloch and Perfit (1981). MK15 is from Makushin volcano on Akutan Island (analyses from McCulloch and Perfit, 1981). ADH is xenolith host rock from Adagdak volcano, Adak Island (DeBari and others, 1987). KAN2 is Tertiary sill on Kanaga Island (Swanson and others, 1987). KAS7A is from Kasatochi volcano (major element analyses from Kay and Kay, 1985a). MOF53A is high-Mg andesite from below Moffett volcano (ADK53 *in* Kay, 1978). Partial analyses and isotopic data included in Kay (1978). New major-element analyses presented here were done by electron microprobe on glasses made from rock powders, and new trace element studies were done by INAA (instrumental neutron activation; Sr and Rb by XRF) at Cornell University. Analytical techniques and standard data are from Rubenstone (1984) and Kay and others (1987). Normalization factors for trace-element plots are Cs (3.77), K (116), Ba (3.77), U (0.015), Th (0.05), Hf (0.22), Ta (0.022), La (0.378), Ce (0.976), Nd (0.716), Sm (0.23), Eu (0.0866), Tb (0.589), Yb (0.249), and Lu (0.0387).

TABLE 2. REPRESENTATIVE ANALYSIS OF HIGH-Al BASALTS FROM THE ALEUTIAN ISLANDS, ALASKA*

Sample	OK7	CHAG	GS727	ADG14	MOF15	MOFA7	KAN508
SiO_2	50.60	51.04	49.68	48.62	49.35	52.40	47.66
TiO_2	1.10	0.98	0.85	0.99	0.86	0.98	1.18
Al_2O_3	18.06	21.12	19.39	18.71	18.57	20.43	18.58
Fe_2O_3	2.99						
FeO	6.75	7.73	9.33	10.27	9.52	8.44	10.28
MnO	0.12	0.12	0.14	0.15	0.15	0.15	0.16
MgO	5.58	3.99	5.60	5.68	4.84	3.52	6.06
CaO	11.46	10.90	10.97	11.52	10.80	9.57	11.41
Na_2O	2.74	3.21	2.66	2.75	2.85	3.27	2.68
K_2O	0.53	0.59	0.82	1.05	1.11	1.39	0.87
P_2O_5	0.28	0.13		0.21	0.15		
Total	100.21	99.81	99.44	99.95	98.20	100.15	98.88
La	5.35	5.88	5.99	8.44	7.76	5.22	7.12
Ce	13.57	13.69	13.86	18.48	18.99	11.28	16.93
Nd		10.76	10.66	11.64	11.67	8.20	11.80
Sm	2.55	2.75	2.77	3.22	3.28	2.09	3.47
Eu	0.929	0.978	0.911	1.040	1.032	0.982	1.085
Tb	0.577	0.519	0.500	0.556	0.469	0.336	0.578
Yb	1.89	1.90	1.91	1.94	1.66	1.40	1.90
Lu	0.280	0.278	0.259	0.270	2.58	0.192	0.266
Ba	191	238	261	441	317	489	352
Cs	0.85	0.40	0.52	0.36	0.32	0.30	0.39
U	0.43	0.29	0.54	0.97	0.72	1.01	0.81
Th	0.88	1.08	1.33	2.08	1.89	2.00	1.55
Hf	1.61	1.58	1.45	1.82	1.78	1.90	1.83
Ta	0.18	0.14	0.19	0.40	0.44	0.00	0.39
Sc	44.2	28.6	35.7	37.6	34.9	19.7	46.5
Cr	72	27	49	41	18	8	35
Ni	9	16	22	18	8	14	18
Co	32	26	32	37	41	16	47
FeO/MgO	1.69	1.94	1.67	1.81	1.97	2.40	1.70
Ba/La	35.7	40.5	43.5	52.3	40.8	93.7	49.4
La/Sm	2.1	2.1	2.2	2.6	2.4	2.5	2.1
La/Yb	2.8	3.1	3.1	4.3	4.7	3.7	3.8
Eu/Eu*	1.01	1.04	0.98	0.96	1.01	1.29	0.96
Th/Hf	0.54	0.68	0.92	1.15	1.06	1.05	0.85
Ba/Hf	119	150	181	243	178	257	192
Ba/Th	218	220	196	212	168	245	227
La/Hf	3.33	3.72	4.15	4.65	4.36	2.75	3.89
Th/U	2.0	3.7	2.5	2.1	2.6	2.0	1.9
$\delta^{18}O$	4.9	6.0			5.8	5.5	

*OK7 is from Okmok volcano, Umnak Island (major-element analyses from Byers, 1961); CHAG is from Chagulak volcano; GS727 is from Great Sitkin volcano (analyses from Neuweld, 1986); ADG-14 is from Adagdak volcano, Adak Island (basalt unit in Coats, 1952); MOF15 and MOFA7 (ADK7 *in* Kay and others, 1978) are from Moffett volcano, Adak Island (main cone, see Coats, 1952); KAN508 is from Round Head, Kanaga Island (Coats, 1952).

tiary sills on Kanaga Island (Pope, 1983; Swanson and others, 1987). Nye and Reid (1986) argue that some of the most Mg-rich (17 to 18 percent MgO, Table 1) of these basalts—those that occur on the flank of Okmok volcano—represent one type of primitive lava. These basalts have only olivine phenocrysts (cores are F092.9), which contain spinel inclusions. Several other Mg-rich basalts (12 to 13 percent MgO) from the same locality have predominantly olivine and minor clinopyroxene phenocrysts (Byers, 1959) and have been proposed as a second type of primitive lava by Nye and Reid (1986). Additional Mg-rich basalts (8.5 to 10 percent Mg), from Okmok and other centers, contain plagioclase as well as olivine and clinopyroxene phenocrysts. Perfit and others (1980a) proposed that a basalt from Makushin (MK15) was parental to the arc suite, and Kay and others (1982) and Conrad and Kay (1984) used one of the Mg-rich Okmok basalts (OK4; Byers, 1961) as a parent in crystal fractionation models that derived more evolved lavas from both the calc-alkaline and tholeiitic centers.

High-Al basalts (17 to 21 percent Al_2O_3) are the most common basalt type in the arc and have been proposed as the primary arc lava by Marsh and coworkers (Marsh, 1981, 1982; Myers and others, 1985, 1986a, b; Brophy and Marsh, 1986; Brophy, 1986). These basalts have high FeO/MgO ratios (>1.8), and low MgO (5 percent), and Ni and Cr content (Table 2); they are characterized by abundant plagioclase (AN80 to 90) and olivine phenocrysts (FO70 to 79) and less common clinopyroxene phenocrysts (Marsh, 1982; S. Kay and Kay, 1985a; Myers and others, 1985; Brophy, 1986). Titanomagnetite is ubiquitous. Although plagioclase-rich, the high Al content of most of these basalts appears not to be related to plagioclase accumulation, since Eu anomalies (Eu/Eu*; Eu* is interpolated from the REE pattern and equals Eu if there is no Eu anomaly) are both positive and negative (Eu/Eu* = 0.85 to 1.1 in 33 samples; Kay, 1977; Myers and others, 1986b; Table 2 and Figs. 4B and 5). Basalts with large positive Eu anomalies (Eu/Eu* > 1.1) and a very high Al_2O_3 content (>21 percent) are much more rare and may have excess plagioclase (Fig. 5). Despite arguments to the contrary by Marsh (1981), plagioclase-rich xenoliths whose minerals are similar to those in high-Al basalts provide evidence that some plagioclase accumulation can occur (S. Kay and Kay, 1985a).

Two end-member proposals have been offered to relate the Mg-rich and high-Al basalts. In one proposal, the Mg-rich lavas are primarily derived from above the subducting plate by melting of mantle peridotite that has been modified by the addition of a component from the plate below (i.e., Kay, 1977, 1980; Perfit and others, 1980a; Kay and others, 1982). In this model, the high-Al basalts are considered to be derivative magmas that form at the crust-mantle boundary or in the lower crust as the result of crystallization of olivine and clinopyroxene, but not plagioclase, from the high-Mg lavas (Conrad and Kay, 1984; S. Kay and Kay, 1985a, b; DeBari and others, 1987; Gust and Perfit, 1987).

In the other proposal, the high-Al basalts result from melting of the eclogitic oceanic crust that is near the top of the down-

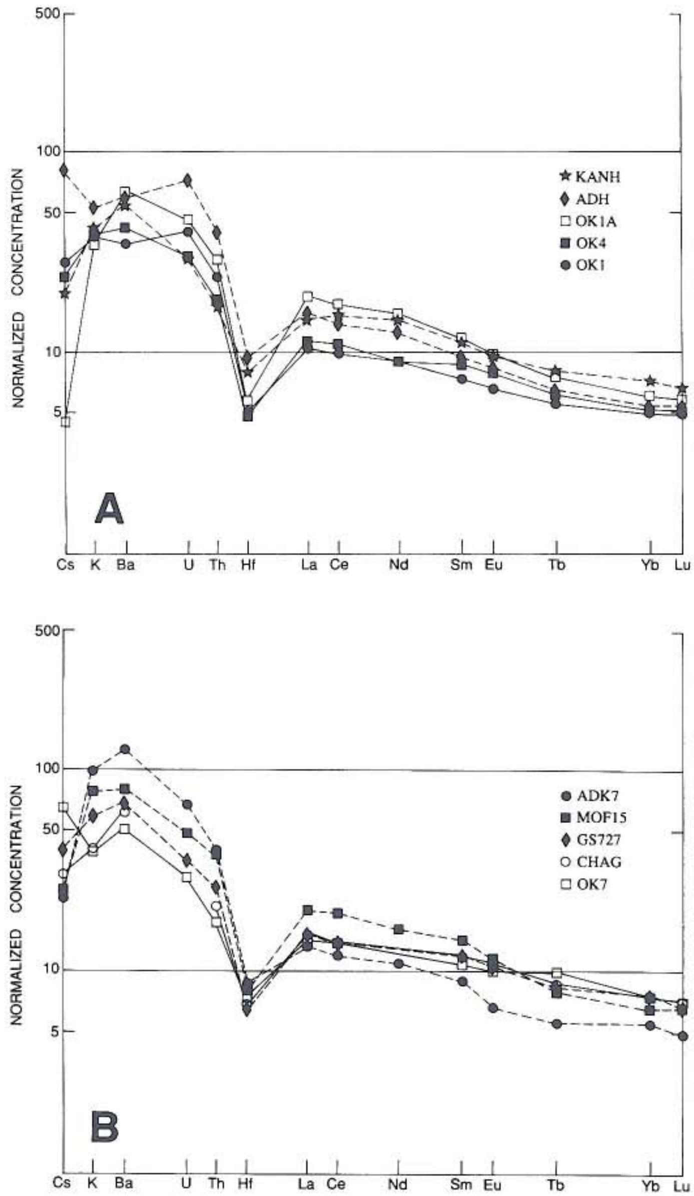


Figure 4. Trace elements of Aleutian basalts (data in Table 1) normalized to chondrites and to ocean-ridge basalts (Cs and K only). See Table 1 for normalization values. A. Mg-rich basalts. OK1, OK1a, OK4 (described in Byers, 1961) from outside the Okmok shield volcano (tholeiitic); ADH is the post-glacial host rock of the ultramafic xenoliths from Adagdak volcano (transitional calc-alkaline; DeBari and others, 1987); and KANH is a Tertiary sill hosting xenoliths on Kanaga Island (Swanson and others, 1987). B. High-Al basalts. ADK7 and MOF15 are from Moffett volcano (calc-alkaline to transitional calc-alkaline); GS727 (data from Neuweld, 1987) is from Great Sitkin volcano (transitional calc-alkaline); CHAG is from Chagulak volcano in the Four Mountains segment; and OK7 is from Okmok caldera (tholeiitic).

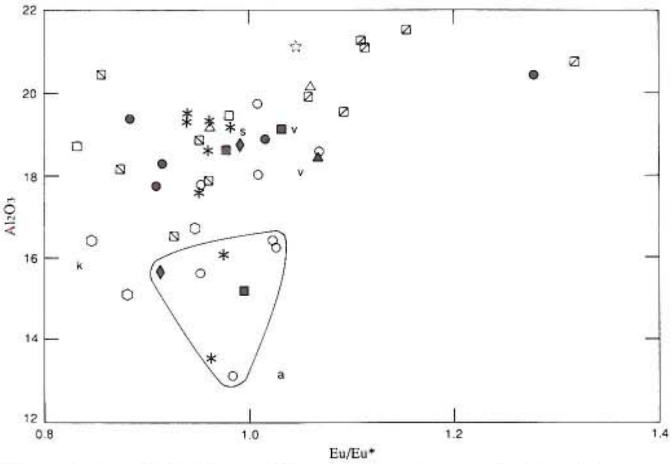


Figure 5. Plot of Al_2O_3 (wt. %) versus the Eu anomaly (Eu/Eu* where Eu* is interpolated from the REE pattern and equals Eu if there is no Eu anomaly) for Aleutian basalts. Symbols are the same as in Figure 2. Open symbols are for tholeiitic and transitional centers (except Great Sitkin, open squares), while closed symbols are for calc-alkaline centers. High-Al basalts with 18 to 20 percent Al_2O_3 commonly have little or no Eu anomaly, whereas basalts with >20 percent Al_2O_3 tend to have positive Eu anomalies. Data are from Kay (1977, 1978), Kay and Kay (unpublished data), McCulloch and Perfit (1981), Myers and others (1986b), and Neuweld (1987). Data from Nye and Reid (1986) are not included because we believe, based on analyses of similar samples from Umnak Island, that there are systematic errors in their Eu anomalies.

going plate. These melts pass through the overlying mantle lithosphere with little or no modification to be parental to Aleutian arc magmas (Marsh, 1982; Myers and others, 1985, 1986a, b; Brophy and Marsh, 1986; Brophy, 1986). In this model, the Mg-rich magmas are considered to be mixtures of high-Al basalts and upper mantle material that has been entrained in them (Myers and others, 1985; Brophy, 1986) or has been assimilated into them (Myers and Marsh, 1987) as they pass through the lithosphere.

The question of whether the Mg-rich lavas are mixtures of xenocrystic mantle material and primary high-Al basalts, or are primarily melts of slab-modified mantle, centers on their isotopic and mineralogic characteristics. Brophy (1986) argued (from a compilation of isotopic data) that the Mg-rich basalts have less radiogenic Sr and more radiogenic Nd than high-Al basalts, which is consistent with contamination of high-Al basalt by MORB-like mantle. However, as Brophy admits, the range of isotopic values is small and the correlation is poor. He supports his conclusion by quoting Myers and others (1985) who suggest that xenoliths described from Moffett volcano by Conrad and Kay (1984) provide evidence of this type of lithospheric contamination. However, these Moffett ultramafic and mafic xenoliths, as well as the dunitic and wehrlitic xenoliths from Adagdak volcano (DeBari and others, 1987) have Nd and Sr isotopic characteristics like those of the arc lavas (R. Kay and others, 1986, see below). Contamination of Moffett lavas by these xenoliths by mixing or assimilation does not change the Nd and Sr isotopic characteristics of the lavas. Furthermore, Nye and Reid (1986) have argued that olivine in the Okmok Mg-rich lavas is in equilibrium with the Mg-rich liquid with which it is associated, and is thus phenocrystic, not xenocrystic. The same is true for olivine in Mg-rich basalts from the Adagdak and Tertiary Mg-rich basalts from Kanaga (S. Kay, unpublished data).

The formation of high-Al basalts from Mg-rich basalts is supported by mass balance models that use observed rock types and are consistent with experimental phase equilibrium studies (Gust and Perfit, 1987). Major-element modeling by Conrad and Kay (1984), Brophy (1986), and Gust and Perfit (1987) shows that fractionation of approximately 20 percent olivine and clinopyroxene from Mg-basalts, such as OK4 and MK15, results in high-Al basalt liquids. The most detailed calculations are those of Conrad and Kay (1984) who combine major-element modeling with the analyses of partitioning of Cr and Ni in zoned clinopyroxene and olivine in cumulate-textured xenoliths to suggest that open-system mixing and fractionation produces a steady-state high-Al basalt. The dunite and wehrlite residuum predicted from these models are observed as xenoliths in the Mg-rich Adagdak basalt (Swanson and others, 1987; DeBari and others, 1987). These xenoliths are isotopically similar to the arc lavas (R. Kay and others, 1986).

Myers and others (1986b) object to these models by arguing that a compilation of trace-element data shows that concentration levels of incompatible elements in high-Al basalts are lower than those in high-Mg basalt, which is opposite to the trend expected if the two are related by crystal fractionation. However, their compilation consists of somewhat evolved tholeiitic basalts (data from Kay, 1977; DeLong and others, 1985) and does not include the high-Mg basalts considered to be parental to high-Al basalt in our discussion. Some of these Mg-rich basalts do have incompatible element concentrations that are lower than the high-Al basalts (Fig. 4A and B). Furthermore, major- and trace-element modeling by Neuweld (1986) shows that some fractionation models are successful: Great Sitkin high-Al basalts (e.g., GS727) can be derived from Mg-basalts like the Okmok sample OK4 (Tables 1 and 2). However, this model is only an approximation, because the parental basalt is from another center (no known Mg-rich basalts occur at Great Sitkin volcano). Because of trace-element differences between and within centers, not all high-Al basalts can be derived from all Mg-rich basalts (see below).

In further support of a derivative origin for Aleutian high-Al basalts, S. Kay and Kay (1985a) have suggested that the phenocryst assemblage common to the high-Al basalts (plagioclase and relatively Mg-, Ni-, and Cr-poor olivine and clinopyroxene) is the result of relatively low-pressure crystallization of high-Al basalt liquids that were produced by high-pressure fractionation of olivine and clinopyroxene in the lower crust. Marsh (1982), Brophy (1986), and Myers and others (1986a) argue that existing experimental evidence does not support this process, but does support the coexistence at equilibrium of anhydrous high-Al basalt liquids and clinopyroxene-garnet-bearing assemblages at 27 kb (Green and Ringwood, 1968; Johnston, 1986) in the subducted plate under the volcanoes. However, the experimental work of Gust and Perfit (1987) on Mg-rich basalt MK15 shows that, at pressures >5 kb, olivine and clinopyroxene crystallize before plagioclase. Interstitial glasses produced in these experiments approach the compositions of high-Al basalts.

Many workers have rejected the hypothesis that arc andesites are primary melts produced by low-percentage melting of garnet-bearing down-going plate, based on models that predict REE that are more fractionated (have high La/Yb ratios) than those in most andesites (see Gill, 1981). However, as pointed out by Apted (1981) and Brophy and Marsh (1986), REE patterns of high-Al basalts (e.g., 37 Aleutian samples have La/Yb ratios = 3 to 5: Table 2, Kay, 1977; Myers and others, 1986a; Brophy and Marsh, 1986) are consistent with a quartz eclogite melting model in which melt segregates at lower pressure and at higher degrees of melting, thereby reducing the amount of garnet (garnet retains the heavy REEs, i.e., Yb) in the solid residual assemblages. In a model of this process, Brophy and Marsh (1986) propose that this low pressure is achieved by decompression of eclogite diapirs. To generate a high-Al basalt (5 percent MgO), Figure 5 of Brophy and Marsh (1986) shows that ~55 percent melting is required if the melt has 5 percent H_2O, and ~65 percent if the melt is dry.

R. Kay and Kay (1985) outline reasons for doubting the validity of the eclogite diapir model. They say that a fundamental question is: at what percentage of melting does the magma segregate from the diapir as it rises after detaching from the subducting plate? Since unmelted eclogite is denser than peridotite, considerable melting is necessary for buoyancy (Brophy and Marsh, 1986, estimate 20 percent). Additional melting occurs during decompression. A key parameter is the segregation velocity at buoyancy (20 percent melt), which Brophy and Marsh (1986) calculate to be 3.2×10^{-4} to 3.2×10^{-8} m/yr, using the porosity-permeability relation of McKenzie (1984), melt viscosities of 10 to 10^5 Pas (10^2 to 10^6 poise), and a grain size of 0.1 mm. This velocity is obviously less than the diapir ascent velocity (presumed to be in the range of 10^{-2} to 10^{-1} m/yr). In contrast, use of the porosity-permeability relation from VonBargen and Waff (1986) and other parameters from McKenzie (1984), including a 1-mm grain size, over the same viscosity range yields segregation velocities of 20 to 2×10^{-3} m/yr. In this alternative calculation, only the most viscous magmas avoid segregation at 20 percent melting. For reference, Crater Lake andesite with 4 percent MgO has a viscosity of slightly less than 10^2 Pas (10^3 poise) at 20 kb and 1,350°C (Kushiro and others, 1976). Experiments on basalt at 10 Kb (Kushiro, 1986) indicate that 1 percent water reduces viscosity by a factor of 3, which is about the effect (in the opposite sense) of a 100°C temperature drop. Therefore, if water is present, equivalent viscosities are reached at lower temperatures. These alternative calculations suggest that, with or without water, andesite should segregate from eclogite at melting percentages below 20 percent, and that high-Al basalt-eclogite equilibrium, which occurs at higher melting percentage, will not be achieved. Note that segregation velocities in clinopyroxene-rich residues will be less than those in the olivine-rich residues assumed in our calculations, although Waff (1986) argues that the effect is small.

To summarize, although melts of silicic composition could be retained in ascending eclogite diapirs at low percentages of melting, melt that approaches high-Al basalt in composition (especially if it were hydrous) would require melt fractions in excess of 50 percent. However, melts segregate quickly at porosities well below 50 percent, halting the diapir. As at mid-oceanic ridges, melt migration by porous flow of basalt through peridotite and finally by magma-fracturing through the overlying lithosphere seems to be a workable alternative. Because this process produces olivine tholeiite (Fujii and Scarfe, 1985), the high-Al basalt is considered to be a derivative melt.

Rare high-Mg andesites (Table 1) from the base of Moffett volcano do have the type of isotopic and chemical signature expected from small degrees of melting of the oceanic crust (Kay, 1978). The steep REE patterns in these rocks suggest that small degrees of partial melt (3 percent melt) could segregate from eclogite in the subducted plate, in contrast to the large degree of melt required for segregation by Brophy and Marsh (1986). These high-Mg andesites further suggest that some magmas may pass through the mantle with relatively little interaction. Although the isotopic data would permit derivation of these andesites ($\epsilon_{Nd} = 11.3$, $^{87}Sr/^{86}Sr = 0.7028$, $^{206}Pb/^{204}Pb = 18.36$; Kay, 1978; R. Kay and others, 1986) from the melting of oceanic basalt buried in the arc crust, their steep REE patterns would be difficult to explain by using such models. Saunders and others (1987) propose that high-Mg andesites from Baja, California, similar to these Aleutian samples, are associated with the eclogite melting that accompanies ridge subduction. However, there is no evidence of ridge subduction contemporaneous with the formation of the Aleutian high-Mg andesites. Their origin remains a problem.

The Mg-rich lavas discussed by Nye and Reid (1986) and Gust and Perfit (1987), raise a further question concerning the composition of the mantle-derived magmas beneath the arc. The question is similar to that posed for ocean-ridge basalts: is the melting product of the mantle a picritic (around 15 percent MgO) basalt that fractionates olivine while still in the mantle, or an olivine tholeiite basalt more like MK15 or OK4 (9 to 10 percent MgO), or depending on conditions, are both types generated? Gust and Perfit (1987) conclude that the lack of three-phase multiple saturation (olivine, clinopyroxene, orthopyroxene) in their experimental studies of olivine tholeiite MK15 shows that MK15 is not a primary magma derived from two-pyroxene peridotite, but must have been derived from a more Mg-rich liquid. Mg-rich magmas have important implications for magma generation in the arc as they have liquidus temperatures above 1,280°C and as high as 1,400°C (Nye and Reid, 1986; Gust and Perfit, 1987).

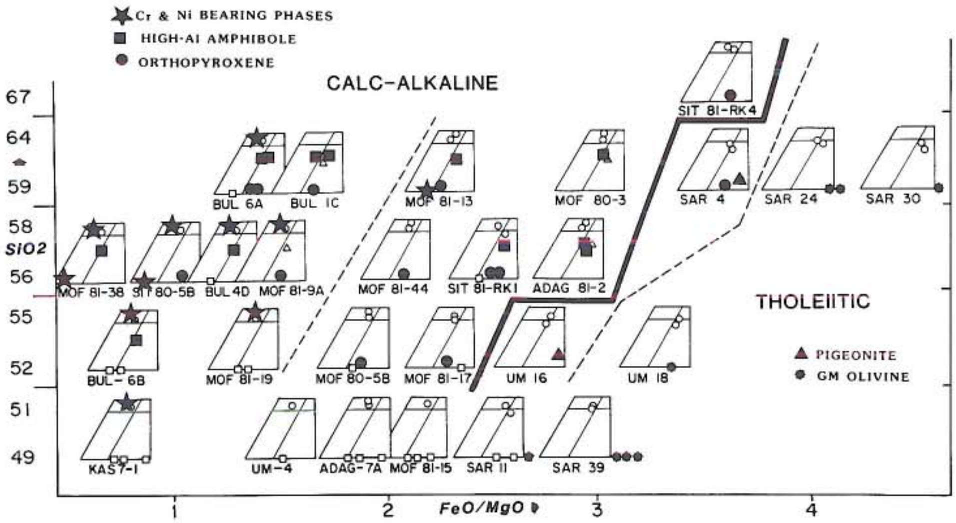


Figure 6. Mafic phases in Aleutian volcanic rocks plotted on the lower half of the Fe* (total Fe as Fe^{+2})-Mg-Ca (molar quantities) triangular diagram (pyroxene quadrilateral). Samples are arranged with respect to their whole-rock SiO_2 content and FeO/MgO ratio. The solid line separates the calc-alkaline and tholeiitic series and dashed lines separate groups within the series. Important mineral phases in distinguishing the series are identified on the figure. Other phases indicated are clinopyroxene (open circles), low-Al amphibole (open triangles) and Ni-poor olivine (open squares). In general, groundmass and phenocryst phases are not distinguished. Symbols represent the phase composition and are based on as many as ten analyses. Adapted from Figure 3 of S. Kay and Kay (1985a).

Mineralogic evidence for the evolution of Aleutian basaltic andesites, andesites, and dacites

The mineralogy of Aleutian volcanic rocks and their xenoliths is an important guide to the crystallization conditions, water content, extent of mixing, and compositions of mixing end members, and thus to the origin of the diversity in Aleutian magmas. Much petrographic information and some optically determined compositions are included in USGS Bulletin 1028 (see references in Kay and others, 1982) and in papers by Coats (1952) and Byers (1961). These studies, along with later microprobe studies by Marsh (1976), S. Kay and Kay (1985a), Brophy (1986), Romick (1982), Neuweld (1987), S. Kay and others (1986b) and Wolf (1987) show that Aleutian lavas are mineralogically diverse (Figs. 6 to 8) and formed under a range of conditions. This mineralogical diversity led S. Kay and Kay, 1985a to suggest that variable amounts of mixing of less fractionated and more fractionated magmas at variable crustal depths can account for much of the magmatic diversity in the arc (Fig. 9). The evidence for these processes is best seen in the basaltic andesites, andesites, and dacites.

Mineral assemblages in the extreme tholeiitic series (i.e., Okmok and Westdahl volcanoes, Figs. 1 and 2) are simple and suggest shallow-level closed-system crystal-fractionation processes (Fig. 9; S. Kay and Kay, 1985a). Evolved basalts (FeO/MgO >1.7) are often porphyritic, but andesites and dacites are generally microphyric to aphyric. Important phenocryst and groundmass phases include olivine, clinopyroxene, and plagioclase (Figs. 6 and 7). Generally, titanomagnetite (Fig. 8) is a groundmass phase. The FO content of olivine and the AN content of plagioclase decrease with increasing whole-rock FeO/MgO ratio and SiO_2 content (Figs. 6 and 7). Olivine (near FO_{50}) is a groundmass phase in the andesites and dacites. The TiO_2 content of titanomagnetite increases with increasing whole-rock TiO_2 content in the basalts and remains relatively constant in the more Si-rich rocks (Fig. 8). As expected in relatively simple fractionation processes, most phases are normally zoned.

Volcanic rocks in the transitional tholeiitic series are more phenocrystic than those in the tholeiitic series, and their mineral assemblages are slightly more complicated. The major mineralogic contrast with the tholeiitic series is that orthopyroxene occurs as a phenocryst in the transitional tholeiitic basaltic andesites, andesites, and dacites, and pigeonite occurs in the groundmass of some transitional tholeiitic andesites (Fig. 6). Normal zoning patterns in plagioclase, olivine, and pyroxene are most common, although some reverse zoning occurs. The mineralogic data for the transitional tholeiitic series are consistent with relatively shallow-level fractionation and some mixing of evolved lavas (Fig. 9; S. Kay and Kay, 1985a). Trace-element studies by DeLong and others (1985) on transitional tholeiitic rocks from Semisopochnoi also suggest that magma mixing occurs in the petrogenesis of the transitional tholeiitic series.

The mineralogy of the calc-alkaline series rocks, which are phenocryst-rich, suggests crystallization at higher pressure and water content and requires more mixing of evolved and unevolved magmas than in the tholeiitic series (Fig. 9). Variations in

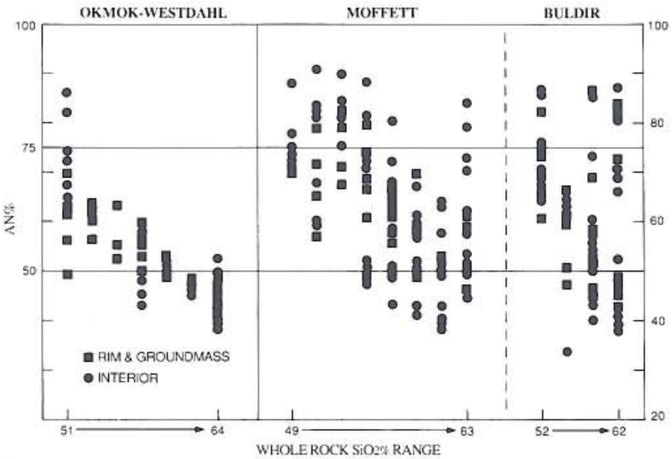


Figure 7. Compositions of Aleutian volcanic plagioclases from two tholeiitic centers (Okmok and Westdahl), a transitional calc-alkaline center (Moffett), and a calc-alkaline center (Buldir) plotted against whole-rock SiO_2 content. AN content of plagioclase (generally microphenocrysts and groundmass phases) from the tholeiitic centers decreases fairly regularly with increasing whole-rock SiO_2 content. In contrast, plagioclases in the calc-alkaline series show a broader composition range whose upper limit in AN content does not correlate with whole-rock SiO_2 content. AN85 (or higher) plagioclase occurs in almost every sample.

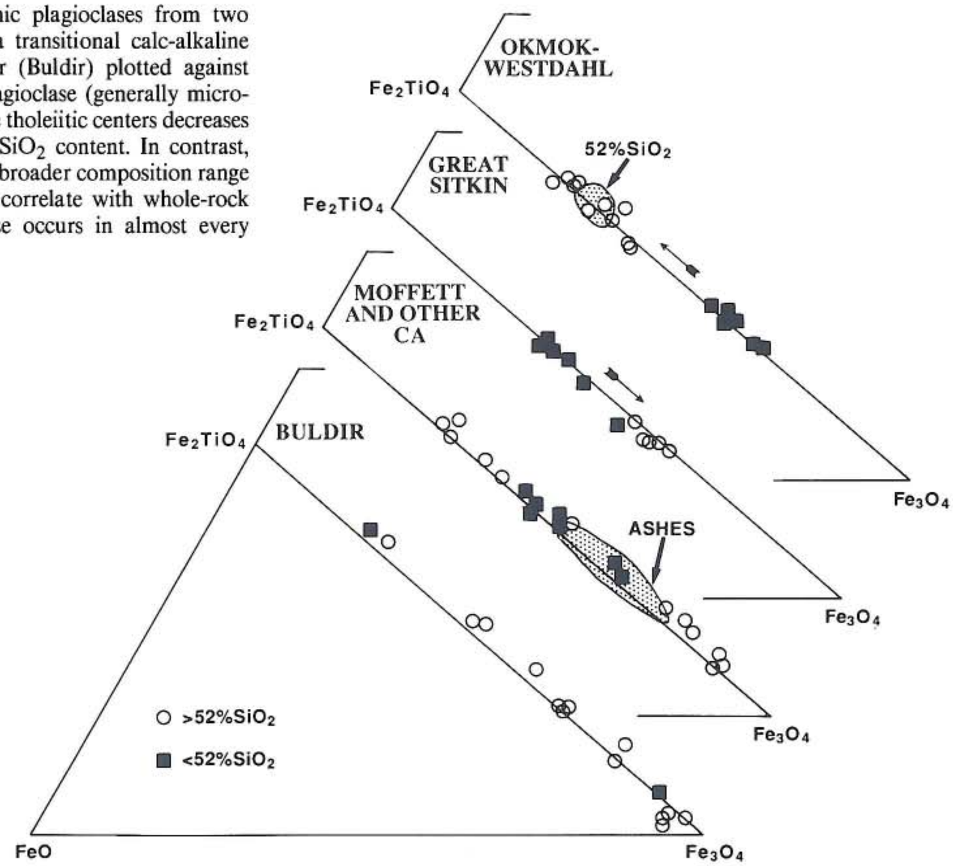


Figure 8. Compositions of magnetites from Aleutian volcanic rocks plotted on the ulvöspinel join (Fe_2TiO_4 - Fe_3O_4) of the system FeO - Fe_2O_3 - TiO_2. Magnetites in volcanic rocks with <52 percent SiO_2 are shown as solid squares, and those in rocks with >52 percent SiO_2 are shown as open circles. In general, magnetite composition correlates with whole-rock TiO_2 content. The magnetites in lavas from tholeiitic volcanoes (Okmok and Westdahl) increase in TiO_2 content as whole-rock SiO_2 increases to 55 percent and TiO_2 increases to 2.7 percent (magnetite compositions in lavas with 52 percent SiO_2 lie in the circled field). In more silicic rocks, TiO_2 content remains relatively constant or decreases as whole-rock TiO_2 decreases. Magnetites in the transitional calc-alkaline Great Sitkin show a decrease in TiO_2 content as whole-rock SiO_2 (50 to 61 percent) increases and TiO_2 (0.85 to 0.65 percent) decreases. Magnetites in the transitional calc-alkaline Moffett lavas are more variable, but the majority follow the same pattern as the Great Sitkin lavas. Most magnetites in calc-alkaline dacites (including ashes) plot in the field labeled ashes. Magnetites in calc-alkaline Buldir lavas are extremely variable, and some are almost pure magnetite. Magnetite compositions are consistent with higher oxygen fugacities in the calc-alkaline series. Data are from S. Kay and J. Romick (unpublished data) and Neuweld (1987).

crystallization conditions and in the composition of mixing end members distinguishes the transitional calc-alkaline and calc-alkaline series rocks (Fig. 9). In the calc-alkaline lavas, the crystallization sequence cannot be determined from plotting percentages of crystals against percentage of groundmass in the manner suggested by Marsh (1981), because magma mixing has resulted in nonequilibrium mineral assemblages.

Volcanic rocks in the transitional calc-alkaline series are characterized by complex phenocryst assemblages with minerals that have complex and often reversed zoning patterns. These characteristics suggest that mixing of high-Al basalts with derivative andesites and dacites is an important process in this series (Fig. 9; S. Kay and Kay, 1985a; Neuweld, 1987; Wolf, 1987). The series is characterized by the presence of either low-Al amphibole (7 to 9 percent Al_2O_3), orthopyroxene, or both as phenocrysts in the andesites and dacites (Fig. 6). Titanomagnetite is an important phenocryst phase (Fig. 8). Both normal and reversed zoning patterns occur in plagioclase, and in clinopyroxene and other mafic phases (Figs. 6 and 7). Compared to the tholeiitic series, the compositions of all mineral phases show more diversity in a single sample and often do not vary regularly with increasing whole-rock SiO_2 content. Plagioclase with $>AN_{80}$ occurs in almost all lavas regardless of their SiO_2 content (Fig. 7). Titanomagnetite compositions are more variable than in the tholeiitic series, and some are high in the magnetite component (Fig. 8). The mineralogy of the andesites and dacites, particularly their amphibole and magnetite phenocrysts, suggests that the transitional calc-alkaline series crystallized under more H_2O-rich conditions at higher fO_2, higher pressure, and lower temperature than the tholeiitic lavas (Fig. 9).

Volcanic rocks of the extreme calc-alkaline series show disequilibrium mineral assemblages and wide compositional variations in complexly zoned minerals. These lavas are modally richer in mafic phenocrysts than other Aleutian magmas (Neuweld, 1987) and are characterized by Ni-rich forsteritic olivine (up to FO_{92} with inclusions of Cr-spinel) and Cr-rich diopside, both of which are frequently partially replaced by high-Al amphibole (Fig. 6). Other phenocryst phases include hypersthene, salitic augite, magnetite, and very complexly zoned plagioclase. Reverse zoning is common in all phenocryst phases. Plagioclase compositions are variable and extend to $>AN_{85}$ over a large range of whole-rock SiO_2 (Fig. 7). Magnetite phenocrysts show a broader range of compositions extending to a higher magnetite component than in other Aleutian magmas (Fig. 8).

S. Kay and Kay (1985a) interpreted the disequilibrium mineral assemblages in the calc-alkaline series as representing mixing of Mg-rich basalt with more evolved andesite or dacite. The lack of plagioclase inclusions in the Mg-rich clinopyroxene and olivine phenocrysts suggests that plagioclase appeared on the liquidus after olivine and clinopyroxene in the mafic mixing end-member (Neuweld, 1987) and supports the crystallization sequence for the Mg-rich basalts discussed above. The replacement of the Mg-rich phases by high-Al amphibole and the occurrence of amphibole phenocrysts in some samples suggests that the mafic end member in these cases had a high H_2O content and that crystallization took place at a relatively high pressure and low temperature within the stability field of amphibole in basalts. These samples also contain salitic augite and Ca-rich plagioclase, which have been correlated with relatively H_2O-rich conditions (Conrad and Kay, 1984). One calc-alkaline andesite (SITR5B) does not contain amphibole phenocrysts but shows convincing evidence for magma mixing: Mg-rich clinopyroxene forms rims on orthopyroxene and occurs as microphenocrysts in the groundmass (S. Kay and Kay, 1985a).

Support for an important role for magma mixing and for the early crystallization of amphibole and titanomagnetite in the evolution of the Aleutian calc-alkaline series also comes from the

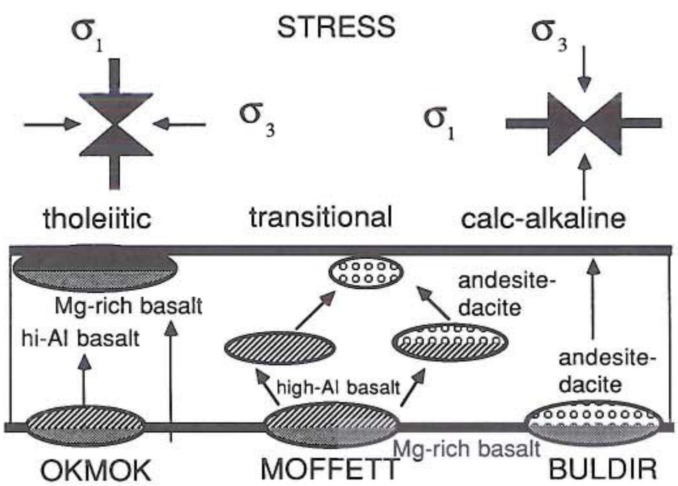


Figure 9. Schematic diagram summarizing differences in depth of fractionation and mixing in the Aleutian volcanic series. Boxed area represents the Aleutian crust and ellipses are magma chambers. In the tholeiitic series (e.g., Okmok), Mg-rich basalts occasionally erupt, but in general they fractionate olivine and clinopyroxene at the crust-mantle boundary, producing high-Al basalts, which then penetrate the crust in a relatively tensional (dike-like) environment. Mafic magmas pond beneath magma chambers causing reheating of partially crystallized magmas and melting of roof rocks. In the transitional tholeiitic (Semisopochnoi) and transitional calc-alkaline series (Moffett), parental Mg-rich magmas also fractionate olivine and clinopyroxene at the crust-mantle boundary producing high-Al basalts. These high-Al basalts can then migrate to magma chambers at various levels higher in the crust, where they can mix with andesites and dacites that have fractionated from high-Al basalts at higher levels, or themselves fractionate to andesite and dacite. Depending on the depth, amphibole may or may not fractionate from these lavas. Melting and assimilation of wall rocks can occur at depth. Because the stress system is mixed, eruption can take place from various levels, but in many cases, final eruption following storage at relatively shallow levels is suggested by low-pressure phenocryst assemblages in the erupted rocks. Calc-alkaline lavas (Buldir) penetrate the crust in relatively compressional (sill-like) environments. Mg-rich basalts fractionate to form high-Al basalts that in turn fractionate to andesites and dacites at depths near the crust-mantle boundary. Mixing of new batches of Mg-rich basalts with these fractionated andesites and dacites occurs at these depths, and the lavas are erupted to the surface relatively quickly as indicated by the presence of amphibole and mineral phases that are not at equilibrium. Remelting and assimilation of older crustal rocks probably alters the final character of these calc-alkaline lavas.

textural characteristics and mineralogy of nondeformed amphibole-bearing cumulate xenoliths found in a pyroclastic andesite from Moffett volcano on Adak Island (Conrad and Kay, 1984). These xenoliths suggest that olivine, clinopyroxene, and in some cases amphibole, crystallized before plagioclase. In particular, plagioclase-free, olivine-hornblende-clinopyroxene rocks (such as MM102; Conrad and Kay, 1984), have subhedral to euhedral Cr-rich amphibole replacing Cr-rich clinopyroxene and Ni-rich olivine (FO_{92}). Conrad and Kay (1984) also showed that mineral proportions of hornblende-bearing gabbro xenoliths are consistent with the residual mineralogy necessary to produce Moffett andesites from high-Al basalt.

The mineralogy of the calc-alkaline lavas and the Moffett xenoliths thus provides evidence for a spectrum of Mg-rich and high-Al basalts as mixing end members in calc-alkaline andesites. Furthermore, the existence of Ni-rich FO_{92} olivine in some of these lavas and xenoliths provides evidence that at least some magmas as mafic as the picritic basalts of Byers (1961) and Nye and Reid (1986) exist beneath both the calc-alkaline and tholeiitic centers.

Major- and trace-element evidence for the evolution of Aleutian basaltic andesites, andesites, and dacites

The mixing and fractionation trends deduced for the various Aleutian magma series from the phenocryst mineralogy are consistent with models based on major- and trace-element chemistry (Coats, 1956, 1959; Kay and others, 1982; S. Kay and Kay, 1985a; Marsh, 1982; Myers and others, 1985; DeLong and others, 1985; Neuweld, 1987; Wolf, 1987). Rare-earth element data are especially useful in supporting the role of amphibole fractionation in the calc-alkaline lavas. In contrast, incompatible trace-element differences (such as U, Th, Ba) do not vary systematically between different magma series and are established in the source region (S. Kay and others, 1986a).

Major-element characteristics of the Aleutian volcanic series are summarized by Kay and others (1982). In general, at the same Si content, tholeiitic series volcanic rocks have higher Fe, Ti, and K and lower Mg, Ca, and Al content than calc-alkaline series volcanic rocks. Differences are minimal in the basalts and increase with increasing Si content. The higher K and Ti content in the tholeiitic centers correlates with higher degrees of fractionation at the same Si content, and the late crystallization of K- and Ti-bearing phases such as titanomagnetite and amphibole. Higher Fe content in the tholeiitic samples can also be partially attributed to the late crystallization of oxide phases. Higher Mg content in the calc-alkaline lavas reflects the greater abundance and higher Mg content of mafic phases, and the higher Ca and Al content correlates with the suppression of plagioclase fractionation and the addition of calcic plagioclase to evolved melts by mixing. Compared with tholeiitic samples, the more primitive character of the major-element compositions (e.g., lower FeO/MgO ratio, higher Mg and Ca content) of the calc-alkaline samples is compatible with the mixing of more mafic lavas with differentiated lavas.

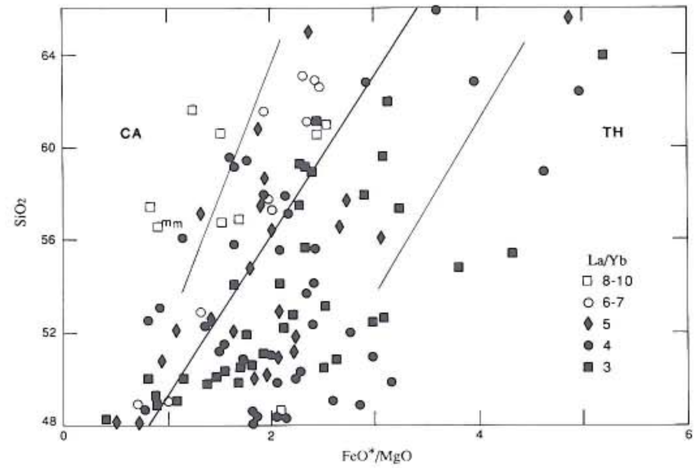


Figure 10. La/Yb ratios of Aleutian volcanic rocks shown superimposed on the plot of SiO_2 content versus FeO*/MgO ratio from Figure 2. The plot shows that lavas in the tholeiitic field and most of the lavas in the calc-alkaline field with <56 percent SiO_2 have La/Yb ratios of 5 or less. On the other hand, most lavas with La/Yb ratios from 6 to 8 and all lavas with La/Yb ratios >8 plot in the calc-alkaline field and have >56 percent SiO_2. The only exception is the Bogoslof basalt (FeO*/MgO = 2) from behind the main arc. All of the rocks with La/Yb ratios >6 contain amphibole, and fractionation of amphibole is an important factor in producing these high ratios. Data sources as in Figure 2.

The REE pattern as indicated by the La/Yb ratios (Fig. 10) is an important trace-element parameter in Aleutian volcanic rocks that correlates with mineral fractionation (Kay, 1977; Kay and others, 1982). As with the major elements, differences are most pronounced in samples with >54 percent SiO_2. As shown in Figure 10, samples with the steepest REE patterns (La/Yb >6) have low FeO/MgO ratios and belong to the calc-alkaline series. In contrast, samples with high FeO/MgO ratios generally have La/Yb ratios <4, while those with intermediate FeO/MgO ratios have La/Yb ratios from 3 to 5. Variations in La/Yb ratio from 3 to 5 among centers can have as much to do with source area differences as with fractionating mineralogy (see below).

The observed differences in La/Yb ratio between Aleutian lavas are consistent with important amphibole fractionation in the evolution of the calc-alkaline, but not the tholeiitic, series. Contrasting trace-element patterns of Aleutian silicic andesites shown in Figure 11 illustrate this point. Amphibole-free tholeiitic andesites like SAR30 have relatively flat REE patterns, with larger Eu anomalies and higher REE levels, than do amphibole-bearing calc-alkaline andesites like MOF3A and BUL6A, which have steeper REE patterns that are more depleted in the middle REE range. REE patterns of low-Al amphibole-bearing calc-alkaline dacitic ashes are even more extreme than these calc-alkaline andesites (Romick and others, 1987). Because amphibole has higher REE distribution coefficients, particularly for the middle and heavy REE, than do olivine, the pyroxenes, the feld-

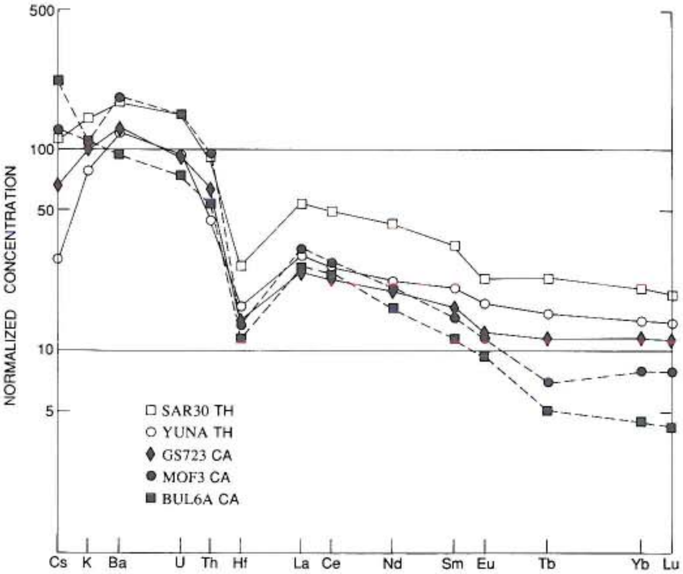


Figure 11. Trace elements for Aleutian silicic andesites normalization to chondrites and ocean-ridge basalts (see Table 1 for normalization values). Samples are from the tholeiitic Westdahl center (SAR30, 62 percent SiO_2, FeO*/MgO = 5.0), the transitional tholeiitic Four Mountains segment center of Yunalaska (YUNA, 63 percent SiO_2, FeO*/MgO = 3.2), the transitional calc-alkaline Great Sitkin center (GS723, 61 percent SiO_2, FeO*/MgO = 2.5), the transitional calc-alkaline Moffett center (MOF3, 63 percent SiO_2, FeO*/MgO = 2.4, data in Table 3), and the calc-alkaline Buldir center (BUL6a, 62 percent SiO_2, FeO*/MgO = 1.3). The tholeiitic dacites have higher REE levels and flatter REE patterns with larger Eu anomalies than the calc-alkaline dacites. The differences are particularly obvious in the levels of the heavy REE. Data are from the Appendix (see microfiche accompanying this volume) and Neuweld (1987).

spars, or the opaques (Arth, 1976), these differences are most easily explained by differences in the relative amount of amphibole fractionation.

The changes in trace-element characteristics in calc-alkaline samples with increasing SiO_2 content also support the idea of amphibole fractionation. For example, high-Al amphibole-bearing samples from Buldir show a progressive increase in the steepness of their REE patterns with increasing whole-rock SiO_2 content (Table 3; Fig. 12A), as would be expected with amphibole fractionation. The relatively constant Hf and Ta concentrations in these Buldir samples may also reflect partitioning of these elements into amphibole. On the other hand, highly incompatible elements, such as U and Th, that are present in very low concentrations in the phenocrysts increase with SiO_2 content, as would be expected with fractionation. The steep REE patterns in these Buldir samples show that the high-Al amphibole in them is not xenocrystic, for if it were, the excess amphibole would cause the REE patterns to become shallower (lower La/Yb ratio), rather than steeper.

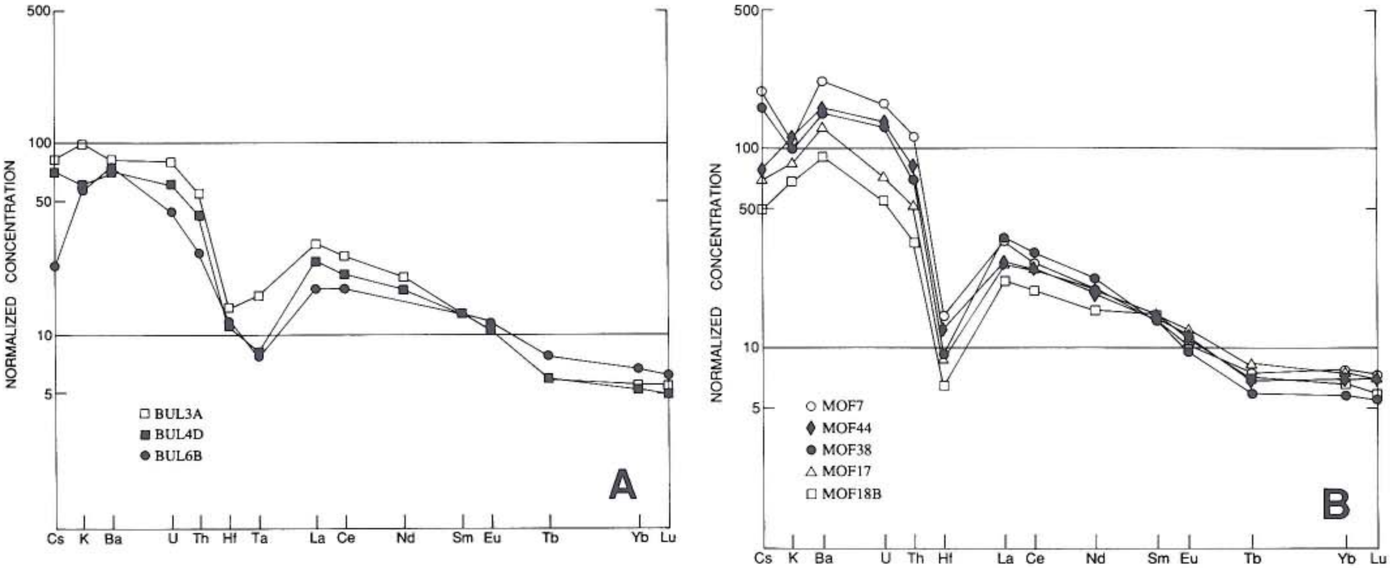
Simple crystal fractionation will not explain the origin of the calc-alkaline lavas. For instance, Moffett andesite MOF38 (Table 3; Fig. 12B), which is similar in many respects to the Buldir samples, cannot be derived from Moffett high-Al basalt MOF15 (Table 1; Fig. 4B) even with amphibole fractionation, because Ni and Cr concentrations in MOF38 are much higher than those predicted by the models. As suggested by the mineralogy and major-element chemistry of calc-alkaline samples like MOF38, their trace-element chemistry also points to mixing of Mg-rich basalts, which are fractionating amphibole with calc-alkaline dacites (S. Kay and Kay, 1985a; Neuweld, 1987). Mixing of Mg-rich basalts and dacites, with no amphibole fractionation, can explain the trace-element characteristics of calc-alkaline andesites like SITR5b (56 percent and SiO_2, FeO/MgO = 1.2), which has a flatter REE pattern (La/Yb = 4.2) than MOF38 and lacks modal amphibole (Neuweld, 1987).

Comparison of the trace-element characteristics of Moffett andesite MOF44 and high-Al basalt MOF15 illustrates another important general point about amphibole in the Aleutian transitional calc-alkaline andesites (Table 3, Figs. 4B and 12B). Unlike the Buldir samples, transitional calc-alkaline andesite MOF44 lacks amphibole phenocrysts and has a REE pattern that is only slightly steeper than that of high-Al basalt MOF15. Although amphibole is not present in either rock, the more concave shape and slightly steeper REE pattern of MOF44 relative to MOF15 and other Moffett high-Al basalts is easiest to explain if some amphibole fractionation occurred. A similar argument has been made by Neuweld (1987) to explain the steeper REE patterns of transitional calc-alkaline andesites relative to high-Al basalts at Great Sitkin volcano. The REE patterns of the Aleutian andesites like these suggest a multistage fractionation process, with amphibole being fractionated before the crystallization of the final phenocryst assemblage.

In contrast to the calc-alkaline series, the trace-element characteristics of tholeiitic series samples are inconsistent with amphibole fractionation and support simple anhydrous crystal fractionation. As shown by the parallel trace-element patterns in Figure 12C for samples from the central region of Okmok volcano (Table 4), trace-element levels including the REE (except Eu which is fractionated by plagioclase) increase with increasing SiO_2 and FeO/MgO ratio. Fractionation models involving plagioclase, clinopyroxene and olivine—phases that have relatively low distribution coefficients for the REE—reproduce these trace-element patterns. The increasing Eu anomaly and lack of increase in the Sr content are consistent with plagioclase fractionation.

Temperature, pressure water content, and oxygen fugacity conditions for the evolution of Aleutian magmatic series

The physical conditions under which the Aleutian magma series formed can be partially constrained by comparing them with experimental analogues. Particularly relevant are the experiments of Baker and Eggler (1983, 1987) on a series of transitional tholeiitic high-Al basalts and andesites from Atka Volcano. The experimental results show that Atka andesites can evolve from Atka basaltic andesites by fractionation of plagioclase, oli-



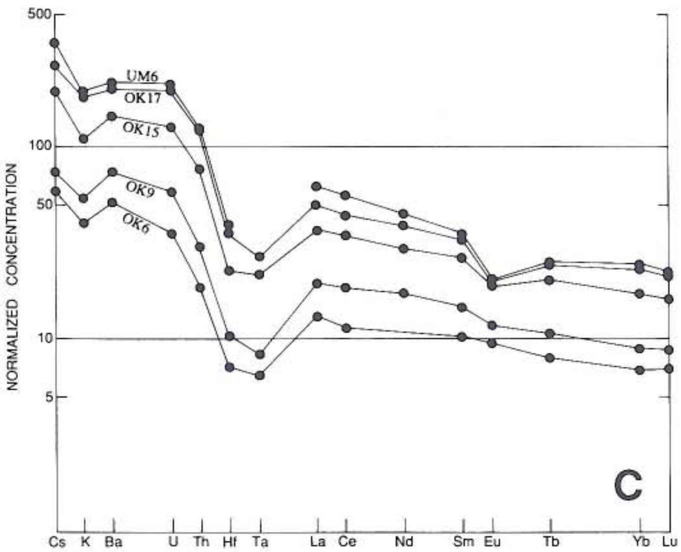


Figure 12. Trace-element patterns of Aleutian volcanic rocks (basalt to dacite) normalized to chondrites and ocean ridge basalts (see Table 1 for normalization values), showing contrasts between the tholeiitic and calc-alkaline series. A. Trace elements for calc-alkaline lavas from Buldir. Samples plotted are basaltic andesite BUL6b (52.7 percent SiO_2, FeO*/MgO = 0.9) and andesites BUL4d (55.9 percent SiO_2, FeO*/MgO = 1.16) and BUL3A (60.8 percent SiO_2, FeO*/MgO = 1.6). Data are in Table 3. B. Trace elements for transitional calc-alkaline and calc-alkaline lavas from Moffett. Samples plotted are high-Al basalt MOF18A (51 percent SiO_2, FeO*/MgO = 2.2), basaltic andesite MOF17 (53 percent SiO_2, FeO*/MgO = 2.1), andesite MOF44 (56.4 percente SiO_2, FeO*/MgO = 1.9), silicic andesite MOF7 (62.8 percent SiO_2, FeO*/MgO = 2.3). Also plotted is calc-alkaline andesite MOF38 (57.3 percent SiO_2, FeO*/MgO = 0.9). Data are from Table 3 and from Appendix (see microfiche accompanying this volume). C. Trace elements for tholeiitic lavas from Okmok. Samples plotted are in order of increasing La: basalt OK6 (50.5 percent SiO_2, FeO*/MgO = 1.8), basaltic andesites OK9 (52.2 percent SiO_2, FeO*/MgO = 1.71) and OK15 (54.8 percent SiO_2, FeO*/MgO = 3.9), and dacites OK17 (63.5 percent SiO_2, FeO*/MgO = 5.2) and UM6 (66.0 percent SiO_2, FeO*/MgO = 5.6). Data are given in Table 4.

vine, and augite between 2 and 5 kb with a magmatic H_2O content of 2 percent, and that Atka dacites can be produced from Atka andesites by fractionation of plagioclase, augite, and orthopyroxene at the same pressure and slightly higher H_2O content. Although complete experimental data are lacking, Baker and Eggler (1987) further predict that Atka basaltic andesites can fractionate from more primitive high-Al basalts under hydrous conditions at 5 to 8 kb.

From these results, Baker and Eggler (1987) conclude that crystallization of Atka lavas is a multi-stage, multi-level process and that the observed phenocryst mineralogy indicates only the final stage. This conclusion is consistent with the multistage crystallization process needed to explain the trace-element chemistry of the transitional calc-alkaline volcanic rocks from Great Sitkin (Neuweld, 1987) and Moffett volcanoes.

Baker and Eggler (1987) have proposed that phase boundaries on the diopside-olivine-quartz and orthoclase (SIOR) pseudoternary projection from plagioclase and magnetite are insensitive to water content, but that pressure expands the olivine phase volume. Projection of compositions of Aleutian samples on this diagram (Fig. 13A) supports the relative crystallization pressures that were suggested by S. Kay and Kay (1985a) on the basis of phenocryst mineralogy. As shown, the projected compositions indicate that the mixing components of the evolved calc-alkaline lavas (Buldir) crystallized at higher pressures than the evolved tholeiitic samples (Okmok anad Westdahl), which plot very near the 1 atmosphere liquid lines of multiple saturation (LLMS). Samples from transitional calc-alkaline to transitional tholeiitic centers (Adagdak, Great Sitkin, and Moffett) plot between these extremes, suggesting that they crystallized at intermediate pres-

TABLE 3. REPRESENTATIVE ANALYSES OF CALC-ALKALINE VOLCANIC ROCKS FROM BULDIR AND MOFFETT VOLCANOS, ALEUTIAN ISLANDS, ALASKA*

Sample	BUL6B	BUL4D	BUL6A	MOF17	MOFA54	MOF38	MOF44	MOF3
SiO_2	52.66	55.90	61.49	52.96	56.49	57.25	57.36	62.88
TiO_2	0.65	0.74	0.49	0.74	0.62	0.44	0.62	0.55
Al_2O_3	18.31	17.57	17.53	18.83	17.27	14.92	17.50	17.49
Fe_2O_3								
FeO	6.78	6.69	4.75	8.68	6.92	6.44	6.91	4.92
MnO	0.11	0.10	0.07	0.13	0.12	0.10	0.11	0.08
MgO	7.35	5.75	3.72	4.19	4.23	7.46	3.58	2.01
CaO	8.78	8.31	6.46	9.94	8.76	8.44	8.42	6.23
Na_2O	3.46	3.18	3.66	3.30	3.37	3.08	3.55	4.18
K_2O	0.81	0.96	1.50	1.17	1.65	1.42	1.57	1.51
P_2O_5	0.14	0.07	0.13	0.22		0.11	0.15	0.16
Total	99.05	99.27	99.80	100.14	99.43	99.66	99.77	100.01
La	6.67	8.99	9.88	10.04	8.99	13.68	10.06	12.07
Ce	17.16	20.54	22.70	24.01	18.68	29.48	24.35	26.93
Nd	7.54	12.40	11.87	14.10	8.23	15.71	13.19	14.13
Sm	2.97	3.04	2.66	3.41	2.66	3.13	3.23	3.31
Eu	1.000	0.956	0.794	1.078	0.881	0.823	0.983	0.994
Tb	0.448	0.365	0.308	0.499	0.349	0.356	0.457	0.421
Yb	1.63	1.34	1.15	1.89	1.25	1.46	2.00	1.97
Lu	0.245	0.194	0.169	0.278	0.196	0.216	0.311	0.302
Rb	18	18	35	28	33		42	49
Sr	377	517	497	572	747		449	474
Ba	256	262	358	474	574	548	581	685
Cs	0.30	0.92	2.84	0.88	0.84	2.08	1.03	1.60
U	0.83	0.82	1.12	1.08	1.80	1.86	1.96	2.22
Th	1.33	2.13	2.77	2.63	3.37	3.96	4.06	4.80
Hf	2.50	2.43	2.53	1.91	2.30	2.04	2.71	2.97
Ta	0.21	0.18	0.18	0.52	0.00	0.28	0.38	0.48
Sc	26.1	30.9	17.8	22.5	23.8	24.8	21.0	10.3
Cr	264	111	86	11	55	447	23	6
Ni	100	36	31	12	24	102	12	2
Co	35	30	55			29	25	20
FeO/MgO	0.94	1.16	1.28	2.08	1.64	0.86	1.93	2.45
Rb/Sr	0.05	0.03	0.07	0.05	0.04		0.09	0.10
Ba/La	38.4	29.1	36.3	47.2	63.8	40.1	57.7	56.8
La/Sm	2.2	3.0	3.7	2.9	3.4	4.4	3.1	3.6
La/Yb	4.1	6.7	8.6	5.3	7.2	9.4	5.0	6.1
Eu/Eu*	1.05	1.08	1.05	1.02	1.11	0.94	1.00	1.00
Th/Hf	0.53	0.88	1.09	1.38	1.47	1.94	1.50	1.62
Ba/Hf	102	108	142	248	250	268	214	231
Ba/Th	193	123	130	180	170	138	143	143
La/Hf	2.67	3.70	3.91	5.25	3.91	6.69	3.71	4.07
Th/U	1.61	2.58	2.46	2.43	1.87	2.13	2.07	2.16
$^{87}Sr/^{86}Sr$					0.70299			
ε_{Nd}					6.2			
$\delta^{18}O$	5.7	7.4		6.1	6.0	3.9	6.0	5.6

*BUL6B, BUL4D, and BUL6A are from Buldir Volcano; MOF17 (parasitic cone, see Coats, 1956); MOFA54, MOF38, MOF44, and MOF3 (main cone, see Coats, 1956) are from Moffett volcano. Major-element analyses largely from Kay and Kay (1985) and isotopic analyses from Kay and others (1978).

TABLE 4. REPRESENTATIVE ANALYSES OF THOLEIITIC LAVAS FROM OKMOK VOLCANO, ALEUTIAN ISLANDS, ALASKA*

Sample	OK6	OK9	OK15	OK16	OK17	OKUM6
SiO_2	50.52	52.23	54.80	57.40	63.50	66.03
TiO_2	0.97	1.24	2.48	1.54	0.81	0.76
Al_2O_3	19.71	17.04	14.90	15.49	16.10	14.81
Fe_2O_3	1.67	3.22	2.88	1.97	1.40	
FeO	6.82	6.60	9.00	7.86	5.00	6.06
MnO	0.12	0.14	0.17	0.15	0.10	0.08
MgO	4.64	5.54	3.00	2.94	1.20	1.08
CaO	11.96	10.39	7.15	6.74	3.80	3.66
Na_2O	2.74	2.90	3.94	4.11	4.50	4.66
K_2O	0.58	0.78	1.23	1.53	2.60	2.62
P_2O_5	0.19	0.19	0.65	0.32	0.23	0.23
Total	99.92	100.27	100.20	100.05	99.24	99.99
La	4.96	7.19	14.40	14.21	18.84	23.65
Ce	11.44	17.63	33.70	33.41	43.82	54.77
Nd	0.00	12.63	22.88	21.24	28.76	32.12
Sm	2.36	3.29	6.69	6.18	7.89	8.10
Eu	0.832	1.037	1.889	1.660	1.660	1.770
Tb	0.483	0.627	1.297	1.210	1.450	1.460
Yb	1.73	2.20	4.21	4.37	5.83	6.29
Lu	0.273	0.337	0.646	0.639	0.851	0.888
Rb						68
Sr						280
Ba	196	284	458	549	780	806
Cs	0.78	0.27	1.91	2.57	4.70	3.52
U	0.55	0.88	1.44	1.90	2.97	3.17
Th	0.92	1.55	2.81	3.84	6.38	6.27
Hf	1.606	2.249	4.180	5.130	7.980	8.320
Ta	0.14	0.18	0.41	0.48	0.61	
Sc	37.0	41.1	40.4	31.1	17.2	17.2
Cr	57	99	5	9	5	
Ni	18	29	2		1	
Co	28	32	25	20	6	
FeO/MgO	1.79	1.71	3.87	3.28	5.22	5.61
Rb/Sr`						0.243
Ba/La	39.4	39.5	31.8	38.6	41.4	34.1
La/Sm	2.1	2.2	2.2	2.3	2.4	2.9
La/Yb	2.9	3.3	3.4	3.3	3.2	3.8
Eu/Eu*	1.00	0.92	0.82	0.79	0.64	0.66
Th/Hf	251	1056	240	214	166	229
Ba/Hf	122	126	110	107	98	97
Ba/Th	213	183	163	143	122	129
La/Hf	3.09	3.20	3.44	2.77	2.36	2.84
Th/U	1.7	1.8	2.0	2.0	2.1	2.0
$\delta^{18}O$	4.2	5.6	4.8	4.2	4.3	5.4

*Samples OK6, OK9, OK15, OK16, and OK17 are from Okmok volcano, Umnak Island (see Byers, 1961, for location and most major elements). OKUM6 is from recent airfall on southern flank of Okmok volcano.

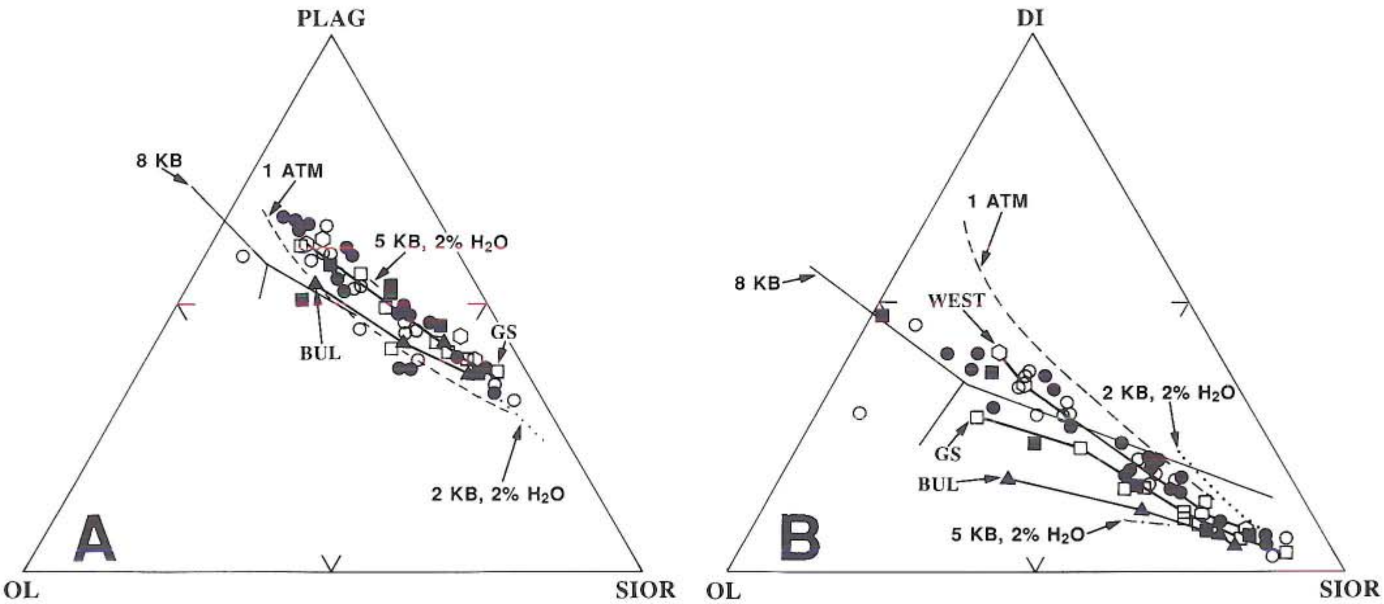


Figure 13. Projections of bulk compositions of Aleutian lavas onto the pseudoternary diagrams of Baker and Eggler (1983). Samples plotted are from the tholeiitic Okmok (open circles) and Westdahl (open hexagons, WEST); transitional calc-alkaline Great-Sitkin (open squares, GS), Moffett (solid circles), and Adagdak (solid squares); and calc-alkaline Buldir (solid triangles, BUL). Solid lines labeled BUL, GS, and WEST are sketched in from the Aleutian data. Labeled lines refer to liquid lines of multiple saturation (LLMS) determined experimentally under the conditions indicated by Baker and Eggler (1983, 1987). A. Plagioclase and magnetite projection onto olivine (OL)-diopside (DI)-silica-orthoclase (SIOR). Pressure moves the LLMS toward the OL corner (both for hydrous and anhydrous conditions). Diagram generally suggests that the tholeiitic centers (WEST) crystallized at lower pressures than the calc-alkaline centers (GS and BUL) in agreement with other criteria discussed in the text. Calc-alkaline samples from Moffett plot surprisingly near the tholeiitic samples. B. Diopside and magnetite projection onto OL-plagioclase (PLAG)-SIOR. Increasing water content shifts the LLMS toward the PLAG corner. The tholeiitic Aleutian data generally plot at lower water content than the calc-alkaline data (GS). Moffett and Buldir samples (BUL) plotting along the anhydrous LLMS are mixtures, and the diagram does not indicate their water content.

sures. Like the Atka transitional tholeiitic samples, nearly all of the transitional calc-alkaline samples plot between the 2 and 5 kb LLMS for 2 percent H_2O.

Baker and Eggler (1987) further propose that phase boundaries on the plagioclase-olivine-SIOR pseudoternary projection from diopside and magnetite are insensitive to pressure, but that increasing water content expands the plagioclase phase volume. In general, the projection of Aleutian samples on this diagram (Fig. 13B) is consistent with tholeiitic samples (Okmok, Westdahl) crystallizing at lower water content than transitional calc-alkaline samples (Great Sitkin, Adagdak, and Moffett). In detail, the tholeiitic samples plot between the 1 atmosphere anhydrous and the 5kb 2 percent H_2O LLMS, and the transitional calc-alkaline samples plot along or above the 5kb 2 percent H_2O LLMS with the Atka samples.

High-Al amphibole-bearing calc-alkaline basaltic andesites (e.g., Buldir) also plot along the anhydrous LLMS in Figure 13B. Their position on this diagram is inconsistent with the higher water content suggested by the occurrence of amphibole in these samples. Instead, the position of these samples could be an artifact—the result of the mixing of Mg-rich basalts and more evolved dacites that were crystallizing under different conditions (S. Kay and Kay, 1985a).

Oxygen fugacity estimates are difficult to make because of the lack of coexisting oxides in almost all Aleutian lavas. However, several lines of reasoning suggest that f_{O_2} is lower in the tholeiitic than the calc-alkaline lavas. For example, titanomagnetite is an important phenocryst in the calc-alkaline lavas, but is rare in mafic tholeiitic lavas and is only present sporadically in the more evolved lavas. The tholeiitic Okmok volcanic lavas have been suggested to form near the fayalite-magnetite-quartz (FMQ) oxygen buffer by Anderson (1976) and Nye and Reid (1986). Anderson's suggestion is based on poorly developed coexisting oxides in an evolved tholeiitic basalt from Okmok, whereas Nye and Reid's suggestion is based on the compositions of spinels included in olivine in Mg-rich lavas from Okmok. In contrast, Brophy (1986) suggests that transitional calc-alkaline Cold Bay lavas are 1 to 2 log units above the nickel-nickel oxide (NNO) oxygen buffer. Brophy's suggestion is based on fugacities calculated from whole-rock compositions based on the method of

Sack and others (1980); it is constrained by the experiments of Baker and Eggler (1983) on the stability of coexisting olivine and magnetite phenocrysts. Additional evidence for f_{O_2} above NNO is suggested by the experimental work of Rutherford and Devine (1986) who found that amphibole was stable in dacitic melts with 4.7 wt% H_2O (H_2O saturated at P = 0.5Gpa) at f_{O_2} of −10.0 (1.5 log units above NNO) at 920°C, but not at a higher f_{O_2}. In addition, Gust and Perfit (1987) show that spinel is a liquidus phase in the Mg-rich basalts from Makushin at fugacities above the NNO buffer but is a subsolidus phase at FMQ.

Differences in f_{O_2} may also be indicated by the generally larger Eu anomalies in the tholeiitic than in the calc-alkaline lavas. As shown in Figures 12C and 14, evolved tholeiitic lavas have moderate to large negative Eu anomalies (Okmok-Westdahl), whereas evolved calc-alkaline lavas (Buldir, Moffett) have small Eu anomalies. Relatively greater decreases in the Sr/Eu* and Eu/Eu* ratios (Eu* is interpolated Eu concentration, based on the concentration of other REE) in the tholeiitic center samples suggest more plagioclase fractionation, which agrees with the larger plagioclase phase volumes indicated in Figure 13B. Some have suggested that the small Eu anomalies in amphibole-bearing rocks occur because amphibole has a negative Eu anomaly that balances the positive Eu anomaly in plagioclase. However, the size of the Eu anomaly depends on the Eu^{+3}/Eu^{+2} ratio in addition to the minerals fractionated; therefore, the anomaly also decreases as conditions become more oxidizing, in agreement with the proposed differences in oxygen fugacities. The slopes (M) in Figure 14 may also suggest that tholeiitic centers (M = 3.6) like MORB (M = 5.0) have f_{O_2} near FMQ. Differences in Sr/Eu* ratios at the same Eu/Eu* suggest regional differences in parental lavas.

The data suggest that amphibole is stabilized in the calc-alkaline centers because of higher pressure and lower temperature crystallization conditions, combined with higher water content and f_{O_2} relative to the tholeiitic centers. The reason that calc-alkaline rocks may have a higher f_{O_2} than the tholeiitic rocks is not known. The differences may reflect differences in mantle history or alteration in the arc crust. Perhaps the mantle is more metasomatized by fluids from the subducted plate in the center of the segments, thus increasing H_2O content and f_{O_2} in the calc-alkaline lavas. The predominance of end-member tholeiitic centers in the eastern part of the arc also may reflect regional differences in mantle (perhaps also crustal) conditions. The f_{O_2} in the calc-alkaline series appears to be higher than that of most oceanic ridge basalts but is similar to that of some oceanic alkaline basalts and included xenoliths (Mattioli and Wood, 1986).

Further evidence for the depths at which Aleutian volcanic rocks evolve; Oxygen isotopes

Preliminary whole-rock oxygen isotope data from fresh Aleutian volcanic rocks (see Appendix on microfiche, at back of this volume) are consistent with chemical and mineralogical arguments for fractionation in shallow-level magma chambers under the tholeiitic centers. Analyses of volcanic rocks over the compositional range basalt to dacite from 15 centers show that while most Aleutian samples are isotopically similar ($\delta^{18}O$ = 5.5 to 6.5) to MORB, the six basalts and andesites with the lowest values ($\delta^{18}O$ = 4.7 to 5.2) are from the tholeiitic centers (Westdahl, Okmok). One explanation for the low values is that magma chamber roof rocks that have been altered at high temperature by

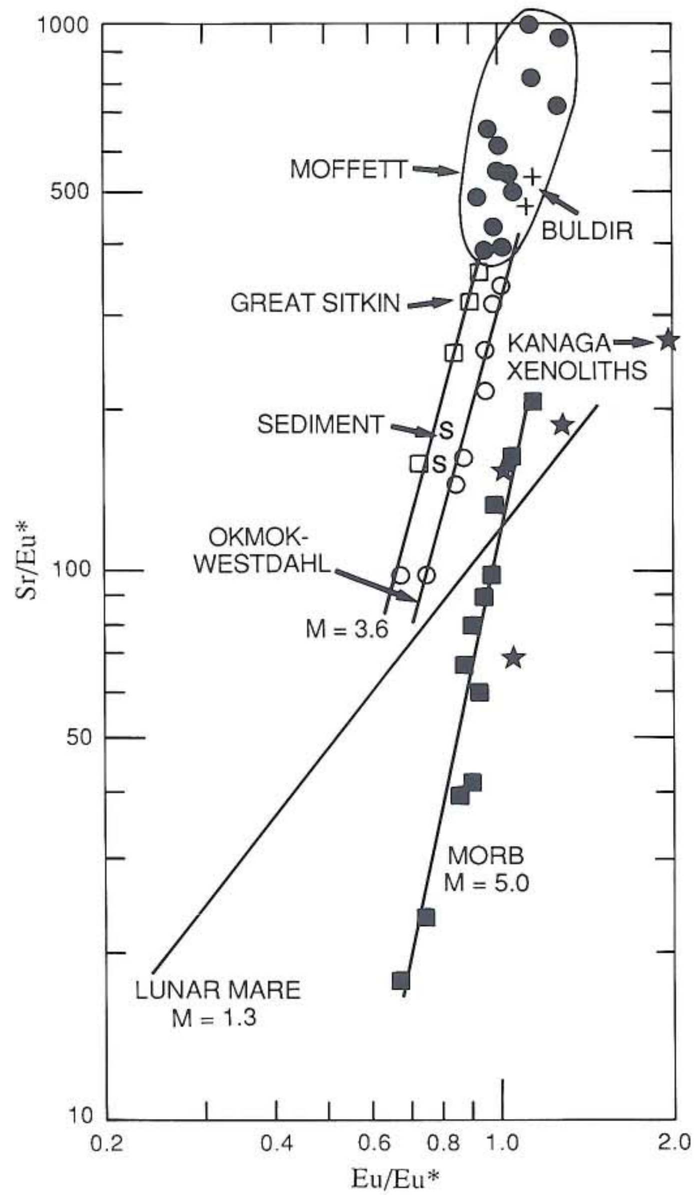


Figure 14. Plot of Sr/Eu* versus Eu/Eu* (Eu* is interpolated from REE pattern) for Aleutian volcanic rocks, xenoliths (from Kanaga Tertiary sill, stars) and sediments (s); MORB (closed squares) and lunar Mare (line represents trend). Slopes (M) of lines through points are suggested to be proportional to the relative fO_2 of the series; that is, Lunar Mare <MORB <Aleutians. The plot also shows the relatively larger Eu anomalies and lower Sr content in the tholeiitic (Okmok-Westdahl, open circles) and extreme transitional calc-alkaline lavas (Great Sitkin, open squares) compared to the more calc-alkaline lavas (circled field; Moffett, solid circles; and Buldir, solid field).

isotopically light meteoric water, founder into the shallow tholeiitic magma chambers underlying the calderas (such as Okmok, 10 km radius), contaminating the magmas with light oxygen (see model of Condomines and others, 1983). The presence of ^{3}He-enriched He in Aleutian hot springs (Poreda and others, 1981) attests to the penetration of light meteoric water into the volcanic pile (and also indicates a high ^{3}He/^{4}He ratio in the volcanic rocks).

Although the oxygen isotope ratios of the Aleutian calc-alkaline series volcanic rocks are not well characterized, 8 of 10 samples from Adak and Buldir volcanoes (δ^{18}O = 5.2 to 6.0) fall within or near the MORB range (δ^{18}O = 5.5 to 6.1). These data are consistent with mid to deep crustal fractionation and mixing processes, and little or no incorporation of isotopically light country rocks from shallow levels into the magmas. Two Buldir andesites have heavier oxygen (δ^{18}O = 7.1, 7.4), suggesting either incorporation of wall rock altered at low temperatures as in some central Aleutian plutonic and volcanic rocks (Perfit and Lawrence, 1979; Kay and others, 1983; Rubenstone, 1984), or more incorporation of oceanic sediment in the source region.

FRACTIONATION-INDEPENDENT TRACE ELEMENTS AND ISOTOPES; THE QUESTION OF SOURCE REGION DIVERSITY IN ALEUTIAN LAVAS

A successful model for the origin of Aleutian arc magmas must identify the mechanism for generating the geochemical characteristics that distinguish Aleutian and other volcanic rocks from the oceanic island basalts (OIB) and mid-ocean ridge basalts (MORB) that occur both north and south of the Aleutian island arc. The model must also explain the diversity of magmatic rocks within the Aleutian arc. The trace-element characteristics that Aleutian magmatic rocks have in common with other arcs include a depletion of some of the high field-strength elements (HFSE; such as Hf, Ta, Ti, Zr), an enrichment of the alkali and alkaline earths (Cs, Rb, K, Ba, Sr), and enrichment of U and Th relative to the light REE elements (Figs. 4, 11, 12A to C, 14, 15A to D). Also, compared to MORB, Aleutian and other arc volcanic rocks are characterized by higher ratios of ^{206}Pb/^{204}Pb, ^{207}Pb/^{204}Pb, ^{208}Pb/^{204}Pb, and ^{87}Sr/^{86}Sr and lower ratios of ^{143}Nd/^{144}Nd (lower ϵ_{Nd}; Figs. 16 and 17). Another important trace-element characteristic of Aleutian and many other arc volcanic rocks (but not island-arc tholeiites, which are absent from the Aleutian arc) is enrichment (relative to chondritic meteorites) in light REE over heavy REE.

Several classes of models have been proposed to explain the geochemical differences between Aleutian lavas and oceanic lavas of the Pacific Ocean crust to the south of the arc. The first class of models involves the addition of sedimentary and hydrated oceanic crustal components from the subducted slab to the source of the parental magmas. In the models of Kay and others (1978), Kay (1980), McCulloch and Perfit (1981), and Perfit and Kay (1986), these components are added from the slab to the overlying peridotitic wedge, which then melts to produce the parental magma. In the models of Brophy and Marsh (1986), Myers and others (1986a), VonDrach and others (1986), and Myers and Marsh (1987), the sedimentary component is melted along with the basaltic oceanic crust in the slab to form the parental high-Al basalts. The high-Al basalts may be contaminated by the assimilation or mixing of material from the lithospheric peridotite wedge and, in extreme cases, become high-Mg basalts. In the second class of models, no sediment is involved, and the parental magma is derived from melting of a mixed MORB- and OIB-type source in the mantle wedge (Morris and Hart, 1983). In this model, a HFSE-bearing residual phase is required to explain the HFSE depletion of the arc magmas.

Examination of the chemistry of Aleutian magmatic rocks shows diversities in incompatible trace elements and isotopic ratios, which seem to suggest both spatial and temporal variability in the source regions of Aleutian arc magmas. Understanding this chemical diversity can help to distinguish and define processes occurring in the mantle wedge and the down-going plate. Although the isotopic ratios (except for oxygen) are independent of fractionation, trace-element ratios may not be; an important requirement in understanding the trace elements is to identify which elements are incompatible and reflect the source region, and which are compatible and controlled by mineral fractionation processes. Thus, U and Th appear to be virtually incompatible, since U-series data suggest that differentiation processes have not changed the Th/U ratio of six historic Aleutian (both calc-alkaline and tholeiitic) lavas by more than about 10 percent (Newman and others, 1984). Determining the compatibility of other elements is not so straightforward (Ryan and Langmuir, 1987), since differences in fractionating mineralogy may reflect differences in source composition, particularly with regard to volatile content and oxygen fugacity.

Trace-element evidence for source region diversity in the Aleutian arc

Of Aleutian magma types, Mg-rich basalts (Table 1; Fig. 4A) have the fewest fractionating liquidus phases (spinel, olivine, clinopyroxene, and in some cases, plagioclase) and should have the largest number of elements that behave incompatibly. Ratios of the most incompatible elements in these basalts are all of arc-type character (see above), but vary in magnitude—an observation most easily explained by source-region diversity. For example, peridotite and gabbro xenolith-bearing basalt from a Tertiary sill on Kanaga Island (KANH) has higher Ta and Hf, and lower Th and U levels relative to La, than other Mg-rich basalts (see Fig. 4A). These characteristics, combined with a relatively low La/Yb ratio (3.2), suggest a less extreme arc signature in this sample. In contrast, a peridotite xenolith-bearing basalt from Adagdak volcano (ADH) has the most extreme arc-type characteristics, since it has the highest Ba, Th, and U and lowest Ta relative to La of any of the Mg-rich basalts. In addition, it has a high La/Yb ratio (4.8). As shown in Figure 4A, trace elements distinguish these two basalts from three Mg-rich basalts from Umnak Island. Of the Umnak group, OK1A is distinct from the other two because of its higher La/Yb ratio (4.8; see Table 1).

High-Al basalts show the same types of incompatible-element ratio variations as the Mg-rich basalts (Table 2; Fig. 4B). Although the high-Al basalts have been involved in more extensive fractionation, mixing, and perhaps assimilation processes than the high-Mg basalts, much of their trace-element diversity is also believed to be due to variations in the source region.

The range and spatial pattern of variations in trace-element ratios in the arc can be illustrated by plotting trace-element ratios for basalts and basaltic andesites (<54 percent SiO_2) against distance along the arc. Representative plots are shown in Figure 15A to D for potentially incompatible element ratios that include the La/Yb (light REE/heavy REE), Ba/La (alkali/light REE), Th/Hf (Th/HFSE), and Ba/Hf (alkaline earth/HFSE) ratios. These plots suggest that while intracenter variations can be substantial, the majority of contemporaneous samples from single centers and centers within a region, have similar ratios, further suggesting regional variation of trace-element characteristics of the magmatic source region. Although the data for many centers are limited and could be argued to be nonrepresentative, a strong case can be made that the more extensive data for Moffett (and Adagdak), Great Sitkin, and Okmok are representative of the range that exists in these volcanoes. An understanding of the chemical diversity in the arc is just beginning, and no model has yet explained the observed intra- and intercenter variations.

Inspection of Figures 15A and B shows that there is little correlation between Ba/La and La/Yb ratios, indicating that the

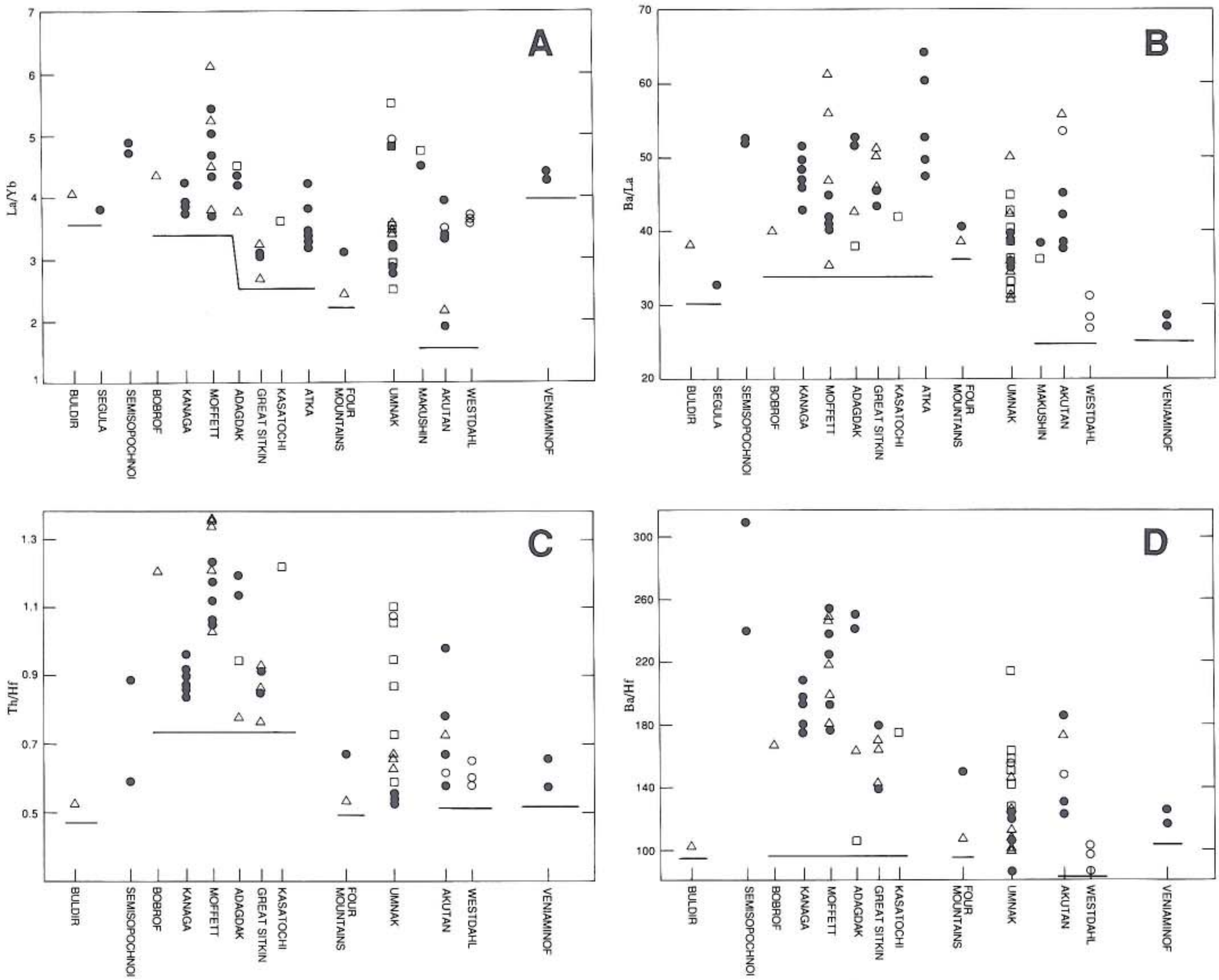


Figure 15. Plots of incompatible element ratios (A, La/Yb, B, Ba/La; C, Th/Hf; D, Ba/Hf) in Aleutian Mg-rich basalts (open squares), high-Al basalts (solid circles), other basalts with <52 percent SiO_2 (open circles), and basaltic andesites (52 to 55 percent SiO_2) (triangles) from centers arranged from west (Buldir) to east (Veniaminof) along the arc. Heavy lines indicate coherent volcanic segments which from west to east area: the Rat, the Andreanof, the Four Mountains, and the Cold Bay segments (see Fig. 1).

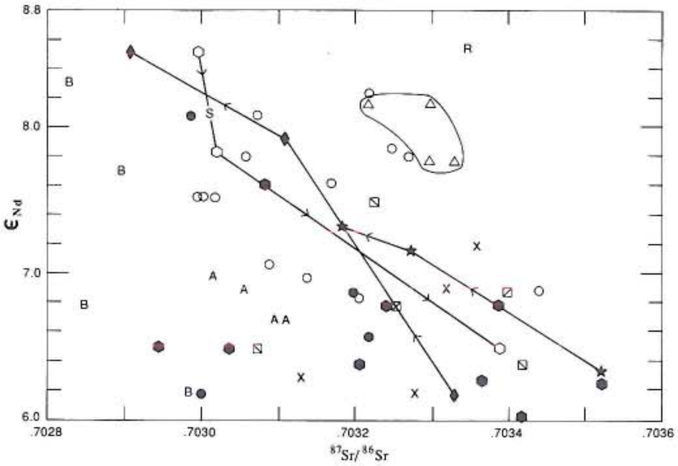


Figure 16. Plot of ϵ_{Nd} versus $^{87}Sr/^{86}Sr$ in Aleutian volcanic rocks. Symbols are the same as in Figure 2 with the addition of samples from Amak (A), Cold Bay (Mt. Simeon and Frosty, filled polygons), and Recheschnoi (R), and xenoliths from Adagdak and Kanaga (X). The Adak Mg-andesite (Kay, 1978) plots off-scale ($^{87}Sr/^{86}Sr = 0.7028$, $\epsilon_{Nd} = 11.3$). Arrows on lines connecting points (diamonds, Makushin; stars, Little Sitkin; hexagons, Westdahl) indicate direction of isotopic change with increasing SiO_2 content. Samples from Semisopochnoi are represented by triangles in circled field. Data are from DeLong and others (1985), Goldstein (personal communication, 1988), R. Kay and others (1978, 1986), McCulloch and Perfit (1981). Morris and Hart (1983), Nye and Reid (1986), VonDrach and others (1986), and White and Patchett (1984). All ϵ_{Nd} data are standardized to BCR = –0.4. $^{87}Sr/^{86}Sr$ ratios are reported as they were in original reference.

REE are decoupled from the alkalis and alkaline earths. This effect can be seen within the Andreanof segment volcanoes (Fig. 1) by noting that Ba/La ratios are in the same range from Atka to Kanaga, while La/Yb ratios are lower at Atka and Great Sitkin than they are at Adagdak and Kanaga. In contrast, both La/Yb and Ba/La ratios are generally lower at Okmok than they are at Moffett. These data provide evidence against the assumption that the range in Ba/La ratios is controlled by mineral fractionation because fractionation of hornblende, the only mineral that can significantly decrease the Ba/La ratio, increases the La/Yb ratio (see trace-element patterns of amphibole-rich xenoliths from Moffett volcano in DeBari and others, 1987). Regional differences in these ratios are thus independent of the chemical and mineralogical criteria that define the Aleutian calc-alkaline and tholeiitic series, and appear to be related to differences in the source regions of the parental lavas (Kay and others, 1986a).

Examination of La/Yb ratios for the arc as a whole shows that samples from both the eastern (Westdahl to Veniaminof) and the western (Buldir to Adagdak) part of the arc have ratios >3.5, whereas many samples from the central arc (Great Sitkin to Akutan) have ratios <3.5 (Fig. 15A: Note that Umnak samples with ratios >3.5 may be older than the main center). Some centers on the Alaska Peninsula that are underlain by continental crust have even higher La/Yb ratios, more like those found in continental-margin arcs (e.g., the Andes; Hickey and others, 1986). It is also worth noting that an important change in the age of the subducting oceanic crust occurs at the Amlia fracture zone (Fig. 1) in the vicinity of an apparent regional break in La/Yb ratios.

Regionally, the Ba/La ratios are lowest in the Alaska Peninsula, generally higher through the central part of the arc, and low again in the west (Segula and Buldir). The lower ratios (20 to 30) in the eastern arc are similar to those in continental arc segments that lack old basement, such as the Andean southern volcanic zone (Hickey and others, 1984, 1986). The Ba/La ratios in the central part of the arc (>40), particularly in the Andreanof segment, are higher than those in most other arcs (Hickey and others, 1984).

Th/Hf (Fig. 15C) and Ba/Hf (Fig. 15D) ratios, which monitor the behavior of Th, U, the alkaline earths, and alkali elements relative to the HFS elements, also show regional variations that correlate with Ba/La. Volcanoes in the Andreanof segment (Kanaga, Moffett, Adagdak, Great Sitkin, and Kasatochi) show the highest Ba/La, Th/Hf, and Ba/Hf ratios and thus have a more extreme arc-type signature than other centers in the arc. Ratios of Ba/Hf are particularly low from Westdahl volcano through the Four Mountains segment.

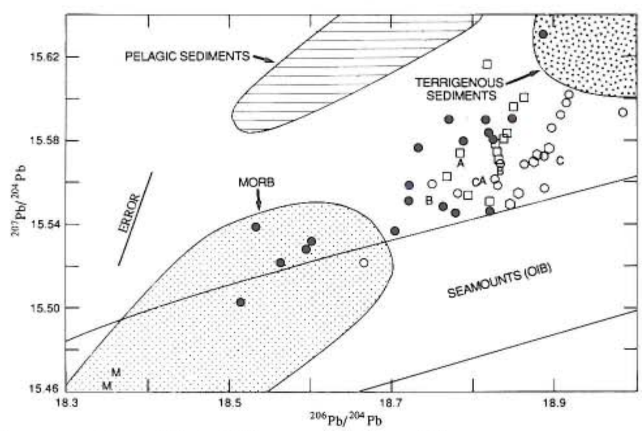


Figure 17. Plot of $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ for Aleutian arc and other northern Pacific volcanic rocks and for Pacific oceanic sediments. Sediments in the pelagic field (horizontal stripes) are from the equatorial and northern Pacific, and those in the terrigenous field (dotted) are northeast Pacific oceanic sediments composed of continental detritus. OIB lavas (field containing seamount data are between parallel lines) are from the northeast Pacific, and MORB lavas (patterned field) are from the Gorda and Juan de Fuca ridges. Data trends can be produced by mixing oceanic type magmas with sediments. Aleutian volcanics from the western Andreanof segment (Adak Island, solid circles; high Mg andesites, M; Atka Island, open squares) have lower $^{206}Pb/^{204}Pb$, and so trend toward pelagic sediments, whereas those from closer to the continental Alaskan Peninsula (Westdahl Volcano, hexagons; Umnak Island, open circles; Cold Bay region, C; Amak Volcano, A; Bogoslof Volcano, B) have higher $^{206}Pb/^{204}Pb$ and trend toward terrigenous sediments. Bar indicates amount and direction of error in the measurements. Sediment and oceanic lava fields are from sources listed in Kay and others (1978). Aleutian data from Kay and others (1978), Morris and Hart (1983), Nye and Reid (1986), and Myers and Marsh (1987). Values are reported as in papers cited.

Although some regional trends emerge, trace-element variations within single centers can be large, suggesting that wide ranges in source components also occur beneath single centers. For example, La/Yb ratios are between 2.5 and 5.5 for most mafic samples along the arc, whereas ratios from Umnak Island samples alone are from 2.8 to 5.0. As pointed out by Nye and Reid (1986), these variations in the Umnak samples can be related to the source region because the Mg-rich basalts have ratios that cover almost this entire range. Looking more closely at the Umnak data, two groups emerge. The first group consists of samples from the main center of Okmok volcano, which have La/Yb ratios from 2.8 to 3.6, and can be interrelated by fractional crystallization (see Fig. 12C). The second group consists of samples distant from the main center that have La/Yb ratios ranging from 2.5 to 5.0 (OK1, OK1A, and OK4 in Fig. 4A). Ages of samples in this second group are poorly constrained, but they probably predate the main center (Byers, 1961), suggesting that part of the variation may depend on time.

To summarize, although some regions are inadequately represented, variations in trace-element ratios that are independent of fractionation imply both regional and local variations in the sources of Aleutian magmas. Volcanoes on the Alaska Peninsula have the most "continental" signatures. Those to the west fall into groups that have La/Yb ratios decoupled from the other ratios. The lowest La/Yb ratios occur in samples from Akutan to Great Sitkin volcanoes, whereas samples from the Andreanof segment have the highest Ba/Th, Ba/La, and Ba/Hf ratios. Before considering the reasons for these variations in more detail, we will discuss the isotopic evidence.

Radiogenic isotopic variations in Aleutian volcanic rocks

The range in the ratios of radiogenic to non-radiogenic isotopes in Aleutian Plio-Pleistocene to Recent volcanic rocks is small, and the ratios are relatively MORB-like compared to many arcs. If the high-Mg andesites of Kay (1978) are excluded, the overall range in ϵ_{Nd} is from +6 to +8.5, in $^{87}Sr/^{86}Sr$ from 0.70295 to 0.70345, in $^{206}Pb/^{204}Pb$ from 18.51 to 19.00, in $^{207}Pb/^{204}Pb$ from 15.50 to 15.63, and in $^{208}Pb/^{204}Pb$ from 38.0 to 38.6 (Figs. 16 and 17; data from Kay and others, 1978; McCulloch and Perfit, 1981; White and Patchett, 1984; Morris and Hart, 1983; Myers and others, 1985; DeLong and others, 1985; VonDrach and others, 1986; Nye and Reid, 1986; R. Kay and others, 1986; Myers and Marsh, 1987; Goldstein, personal communication, 1988).

As with incompatible trace elements, the range of isotopic ratios in the basalts is almost as great as that of samples covering the entire compositional range. For example, we note the isotopic differences among Mg-rich basalts from Umnak (OK1A, ID1, ID16, ID18, ID25) and Makushin (MK15) and high-Al basalts from Adagdak (AD14), Semisopochnoi, Akutan, and Umnak (Table 1 and 2; Figs. 16 and 17). As Nye and Reid (1986) point out, the Umnak samples show that considerable variation can occur in a restricted geographic region. However, since the ages of the older of these samples are unknown, the extent of variation with time is uncertain. The data show that the cause of variation is not simply the result of crustal contamination (which should be minimal in the Mg-rich basalts). Thus, inhomogeneities resulting from subcrustal processes must be instrumental in controlling these variations.

Nd and Sr isotopic ratios for all analyzed Aleutian lavas and xenoliths show only a rough correlation (Fig. 16) and are not easy to interpret. As pointed out by VonDrach and others (1986), samples from the secondary front islands (Amak and Bogoslof) plot below the trend of main arc. Some amphibole-bearing xenoliths and lavas from both Adak and Cold Bay volcanic centers (calc-alkaline centers on the main front) also plot below the trend of the main arc. Isotopic variations in these Moffett and Cold Bay samples show no simple correlation with major- or trace-element chemistry. All samples from Semisopochnoi and four samples from Umnak Island plot above the main trend. Samples that plot in the main Aleutian trend vary inconsistently with SiO_2 content. For example, three samples from calc-alkaline Makushin volcano show a decrease in $^{87}Sr/^{86}Sr$ and an increase in ϵ_{Nd} with increasing SiO_2 (McCulloch and Perfit, 1981), as do three samples from calc-alkaline Little Sitkin (White and Patchett, 1984). In contrast, the reverse pattern can be found with increasing SiO_2 in samples from tholeiitic Okmok and Westdahl (Goldstein, personal communication, 1988). At the transitional-tholeiitic Atka volcanoes center, samples with relatively small variations in ϵ_{Nd} (VonDrach and others, 1986) also show an increase in $^{87}Sr/^{86}Sr$ (0.70319 to 0.70345, in Myers and others, 1985) with increasing SiO_2 (Frost and Myers, 1987).

Regionally, isotopic and trace-element data show some correlations. VonDrach and others (1986) noticed that ϵ_{Nd} is lower in the Andreanof segment and on the Alaska Peninsula than it is in the western arc volcanoes and in the Okmok region. Although exceptions occur when the data of Goldstein (personal communication, 1988) are added, the pattern generally holds (Fig. 16). The generally lower ϵ_{Nd} in the volcanic rocks of the Andreanof segment correlates with their greater enrichment in the alkali and alkaline earth elements. However, lower ϵ_{Nd} and the trace-element enrichment-depletion patterns observed in the Cold Bay and Rat segments do not correlate, nor is a one-to-one correlation found in samples of the Andreanof segment.

Within the Andreanof region, Myers and others (1985) have suggested that volcanic samples from Atka show a smaller range in Sr isotopic ratios than samples from the Adak volcanoes. Additional data from Adak (R. Kay and others, 1986; Goldstein, personal communication, 1988) fall in the same range. Excluding a Mg-rich andesite from Adak (Kay, 1978), ϵ_{Nd} data ranges from 7.2 to 8.3 on Adak and from 6.7 to 8.0 on Atka (value of BCR standard not given by Frost and Myers, 1987).

Myers and others (1985) have used the range of Sr isotopic values within a center to support their contention that centers such as Atka represent "clean" (mature) systems, whereas centers such as Moffett represent "dirty" (immature) systems. In this model, chemical diversity occurs in the immature (calc-alkaline) centers because lithospheric mantle debris is picked up by the magma. In contrast, the mature (tholeiitic) centers have well-established magma conduits that are protected from lithospheric debris by thermal-chemical boundary layers. Although the model could explain the lower Sr isotopic values of some of the Moffett samples, samples like ADK54 which has both low Sr and low Nd isotopic ratios (Goldstein, personal communication, 1988) are difficult to explain by lithospheric debris.

In a somewhat similar model, the apparently greater isotopic variability in the calc-alkaline centers than in the tholeiitic centers—like their greater whole-rock chemical and mineralogic diversity (Kay and others, 1982)—can be explained by magma mixing. It is a plausible assumption that each mantle-derived magma batch is isotopically distinct because it has different amounts of sediment incorporated and mantle inhomogeneities can be variably modified by crustal contamination. The inefficient mixing of compositionally heterogenous magmas combined with assimilation and fractional crystallization (AFC) at intermediate to deep crustal levels in the calc-alkaline centers would result in some homogenization of incompatible element and isotopic signatures, but each mixed lava would retain evidence of its previous history. In contrast, in higher volume tholeiitic centers where larger magma batches more easily reach higher level magma chambers, isotopic signatures can be more uniform.

Sediment subduction in the Aleutian arc; Evidence from Pb and O isotopes and ^{10}Be data

Both of the models discussed above assume that subducted sediment plays a role in shaping the geochemical character of the source of Aleutian lavas. The available Pb isotopic data can be used to support the idea of sediment contamination in Aleutian lavas (Kay and others, 1978; Sun, 1980). On a plot of $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ (Fig. 17), the Aleutian data fall in a field that diverges from the regional MORB and OIB fields toward the field of oceanic sediments (Kay and others, 1978; Nye and Reid, 1986). The Andreanof samples (Adak and Atka) have lower $^{206}Pb/^{204}Pb$ ratios suggesting a slightly larger pelagic sediment component than samples from the eastern arc (Umnak, Westdahl, Cold Bay). Pb isotopic values from Amak and Bogoslof, volcanoes behind the main arc, are indistinguishable from the other Aleutian samples.

In an alternative explanation, Morris and Hart (1983) have argued that the Aleutian Pb data lie within the global oceanic island basalt (OIB) field and could have an OIB mantle source. However, Perfit and Kay (1986) have pointed out that most of the Aleutian Pb values do not overlap the OIB field if the ^{207}Pb-rich Southern Hemisphere (DUPAL—type) OIBs are excluded. For instance, the least radiogenic of the Umnak Mg basalts (ID1, ID16) fall close to the Northern Hemisphere seamount (OIB) field and can be explained by an OIB mantle source without another component (Nye and Reid, 1986, their Fig. 5). But Pb data on other Aleutian samples are difficult to explain without a contribution from a source with more radiogenic ^{207}Pb.

^{10}Be data from young Aleutian volcanic rocks (Tera and others, 1986) also strongly suggest a sediment component (all rocks measured have ^{10}Be well above detection limits). As a cosmic-ray-produced nuclide with a 1.5-m.y. half-life, ^{10}Be is present only at the Earth's surface or at the top of oceanic sediment columns. Thus, ^{10}Be in the Aleutian volcanic rocks must be derived from subducted sediments or contamination at the Earth's surface, because ^{10}Be that was originally present in rocks buried in the crust of the main Aleutian arc would have completely decayed. The positive correlation of ^{10}Be and other incompatible elements in the Bogoslof 1927 basaltic and 1796 andesitic eruptions suggests that ^{10}Be is not the result of surface contamination. Recent work by Monaghan and others (1988) supports this suggestion because both whole rock and mineral separates from two Bogoslof eruptions have the same $^{10}Be/^{9}Be$ ratio, showing that the ^{10}Be was incorporated into the magma before the phenocrysts crystallized. We infer crystallization to be relatively deep, because the Bogoslof rocks are amphibole bearing. The high ^{10}Be content of the Bogoslof samples, which come from behind the main arc (Fig. 1), suggest that some sediment component reaches a depth of at least 150 km in the subducting slab (Morris and others, 1989).

The general correlation between the ^{10}Be and Ba content in volcanoes from both the eastern and central Aleutians (Fig. 18) suggests that enrichments in Ba, like those in ^{10}Be, are at least partially related to a subducted sedimentary component. Although the data are extremely limited, the somewhat higher ^{10}Be/Ba ratios in flows from the eastern volcanoes may correlate with the generally higher $^{206}Pb/^{204}Pb$ (Fig. 17) and lower Ba/La ratios (Fig. 15b) in the volcanic rocks of this region, suggesting that there are regional differences in the character of the sedimentary component being subducted along the arc.

Although a sedimentary component appears to be present in the source region of Aleutian volcanic rocks, how this component is mixed into the magma source remains a problem. Because the compositional variability within the sedimentary column is extreme (especially for trace elements) compared with the variability of the volcanic rocks, large-scale mixing is required (VonDrach and others, 1986; Kay and Kay, 1988). Calculations based on trace-element and isotopic data by Kay (1980, 1984) suggest that the amount of sediment relative to the total mass of mantle peridotite source is less than 1 percent. Even at this low percentage, a substantial fraction of elements such as Pb and Ba are furnished by the sediment. This model contrasts with the extreme view of Myers and others (1986b) that 8 to 13 percent carbonate and 21 to 44 percent of pelagic sediment are added to an eclogite melt to create parental high-Al basalt magmas. The high percentage of carbonate in this model is difficult to reconcile with the observed $^{87}Sr/^{86}Sr$ in the Aleutian arc lavas, and if applied to other arcs with higher $^{87}Sr/^{86}Sr$, suggests extremely high percentages of carbonate in those arcs.

The oxygen isotopic composition of Aleutian lavas also argues against the large quantities of sediment in the model proposed by Myers and others (1986a). Nye and Reid (1986) observed that $\delta^{18}O$ values in Umnak lavas ranged from 5.2 to 6.5, which is within the range of mantle-derived tholeiitic rocks on

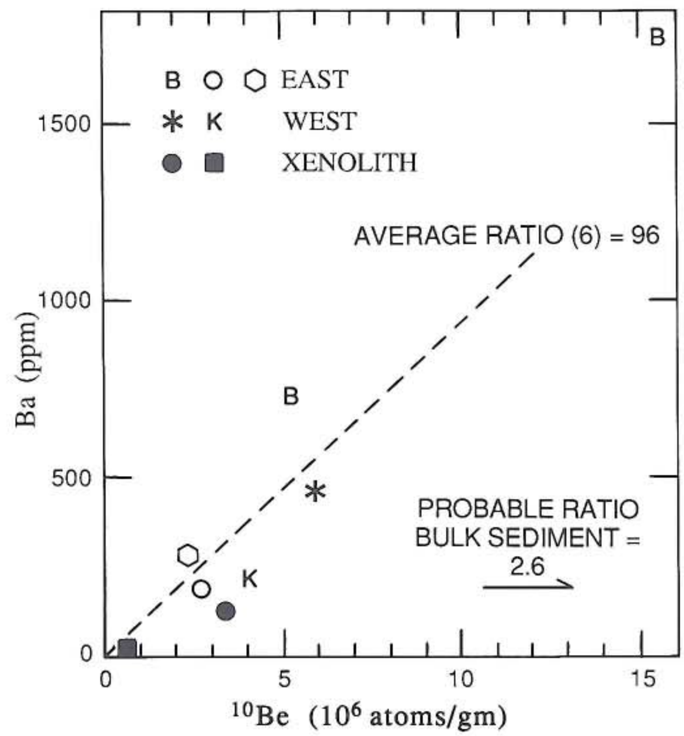


Figure 18. Ba versus ^{10}Be content for recent Aleutian volcanic rocks from Kanaga, Kasatochi, Okmok, Bogoslof, and Westdahl volcanoes (symbols are the same as in Fig. 2). Also plotted are ultramafic xenoliths from Moffett and Adagdak volcanoes. ^{10}Be data are from Tera and others (1986). Probable ratio in units of the figure) of the bulk sediment secton south of the Aleutian trench is estimated to be 2.6 by Tera and others (1986).

Hawaii (4.9 to 6.0) and in MORB (5.8 ± 0.3) as noted by Kyser and others (1981). A survey of lavas from all along the arc (see Tables 1–5 and Appendix in microfiche) shows that almost all samples, ranging from basalts to dacites. fall in only a slightly larger range ($\delta^{18}O$ = 4.8 to 6.5; 64 samples). Simple mass balance shows that only a small proportion of oxygen from isotopically heavy sediments ($\delta^{18}O$ >+20) can exist in these Aleutian lavas. Although more work is needed, it appears that primary Aleutian olivine tholeiites may have lower $\delta^{18}O$ than MORB (by ~one-half unit); it thus remains to explain light rather than heavy oxygen. The $\delta^{18}O$ data are consistent with the model of Morris and Hart (1983), because OIB have lower $\delta^{18}O$ than MORB (Kyser and others, 1981), but that model alone cannot explain the ^{10}Be data.

Some of the isotopic heterogeneity in Aleutian volcanic rocks could be generated as the lavas traverse the basal arc crust, as well as in the source region, as postulated for some Aleutian plutons (R. Kay and others, 1986; Kay and others, 1990). However, it is evident from the variable isotopic and trace-element signatures within single volcanoes, and especially within the most mafic arc magmas, that the mantle source beneath the arc is inhomogeneous.

GEOGRAPHIC AND TECTONIC CORRELATIONS

A goal of Aleutian studies is to correlate regional chemical and isotopic variability with regional tectonics. As a control on magmatic evolution in the crust, Kay and others (1982) postulated that tholeiitic magmas evolve in shallow magma chambers in regions of the arc that are under extension, and that calc-alkaline magmas evolve in deeper magma chambers in regions of the arc that are under compression. Regional tectonics exerts control on magma formation in two other ways: (1) by determining source and lithospheric mineralogy, and (2) by determining the proportion of components in the source. The basis for these controls can be outlined, but the details of their applicability remain obscure.

Mineralogic differences in the mantle source region could be related to the age of the oceanic crust that is being subducted. The intersection of the Amlia fracture zone with the arc (Fig. 1) between the Andreanof and Four Mountains segments is a case in point. The subducted crust to the east of this zone (Four Mountains side) is older than the subducted crust to the west (Andreanof side). Although the seismic data are not definitive (K. Jacob, oral communication, 1987), the older age of the subducting crust to the east (Four Mountains side) suggests that the slab would be thicker and heavier, and thus could be subducting at a greater angle in the east than to the west. A higher subduction angle would result in a thicker and hotter asthenosphere over the subducted plate, leading to more spinel peridotite relative to garnet peridotite in the wedge and higher degrees of melting in the peridotite source. Either less garnet in the source region or higher degrees of melting would lead to less retention of heavy REE in garnet and lower La/Yb ratios. Trace elements not fractionated by garnet are unaffected. However, the regional change in La/Yb ratio does not coincide exactly with the Amlia fracture zone, and the model does not explain why La/Yb ratios are high in the Rat segment, where the subducted oceanic crust is also younger.

Speculating further, differences in the subduction angle might also explain why there are relatively fewer calc-alkaline rocks in the region east of the Amlia fracture zone. If the peridotite source overlying the mantle is hotter in the east, the percentage of melting and thus the magma production rate are higher, leading to large shallow magma chambers. To the west the mantle is cooler, leading to a lower magma production rate and a greater chance that magmas will crystallize in the crust rather than erupt. If water is released from the subducted plate at equal rates east and west of the Amlia fracture zone, higher water content in the calc-alkaline centers (e.g., Moffett) might simply reflect the lower melting percentages and lower regional magma production rates.

Similarly, for secondary arc centers such as Bogoslof (north of Okmok), low production rates and higher water content (magmas are amphibole-bearing) correspond to a thicker lithosphere under the volcano. Thus, the higher K_2O (and trace element) content in the Bogoslof lavas could reflect less partial melting in the source.

Regional differences in trace-element content and isotopic ratios of arc magmas are related to differences in the proportions of the components that make up the magmas. Identification of the exact components remains a problem (except for ^{10}Be, which is a tracer of top sediments). The trace elements and isotopes that represent geochemically coherent "excess components" (e.g., non-OIB or MORB, see Kay and Kay, 1988) are a valuable tracing tool, and provide a constraint on the cause of regional differences in the arc.

A promising hypothesis to explain the regional differences in the arc involves differences in sediments that are subducted at the trench. Modern lavas in the Andreanof segment tend to have higher concentrations of Ba, Th, U, Rb, and Sr relative to La, higher $^{10}Be/Ba$ ratios, lower ϵ_{Nd}, and lower $^{206}Pb/^{204}Pb$ ratios than lavas from the Four Mountains segment and the adjacent region to the east (e.g., Umnak Island). Since the Andreanof segment is in the central part of the arc, far from both the Aleutian abyssal plain to the east and the Meiji sediment tongue to the west (Vallier and others, this volume), the subducting sediments might be more pelagic and less terrigenous. The closer trend toward the pelagic sediment field of the Andreanof segment samples on the Pb isotope diagram in Figure 17 supports this hypothesis. Proof, however, requires more data on the chemical character and distribution of sedimentary types in the Aleutian arc region. Interestingly, the boundary between the Rat and Andreanof segments corresponds to one of the trench sedimentation changes noted by Mogi (1969), which might explain the change in the range of Ba/La, Ba/Hf, and Th/Hf ratios that occur there.

A second hypothesis to explain the chemical variations in magmatic rocks along the arc involves the recycling of older Aleutian crust. This older crust includes both oceanic and older arc crust (S. Kay and Kay, 1985b) in the oceanic part of the arc, as well as continental crust in the Alaska Peninsula. Broadly speaking, these crustal differences are not reflected in the isotopic data (e.g., Kay and others, 1978; VonDrach and others, 1986), suggesting that crustal contamination is a second-order effect. Nonetheless, since the arc crust is relatively young, involvement of the preexisting crust is difficult to determine from isotopic data, and the extent of intracrustal recycling remains an important question. Intracenter isotopic variations that correlate with the stage of crustal differentiation (e.g., volcanic trends, Fig. 16; pluton trends, R. Kay and others, 1986) are probably due to crustal-level processes involving assimilation and mixing with older rocks. Trends toward higher ϵ_{Nd} in more silicic rocks could result from melting or assimilation of old oceanic crust buried in the arc. However, some important temporal chemical changes in Adak Island magmatic rocks described in the next section cannot easily be attributed to recycling of the preexisting crust.

Clear chemical differences exist between Tertiary arc-type basement rocks on Adak Island in the central Aleutians and spreading center–like basement rocks on Attu Island (173°E) in the volcanically inactive western Aleutians (Fig. 19; Table 5; Rubenstone, 1984; R. Kay and others, 1986). Since little work has been done on basement rocks in the region between the two islands, the effect of these differences on modern volcanic rocks is hard to assess. One logical place for the basement to change character is at the intersection of Bowers ridge with the arc—a change that could explain some of the trace-element differences between the rocks west and east of the western end of the Andreanof segment (Fig. 15).

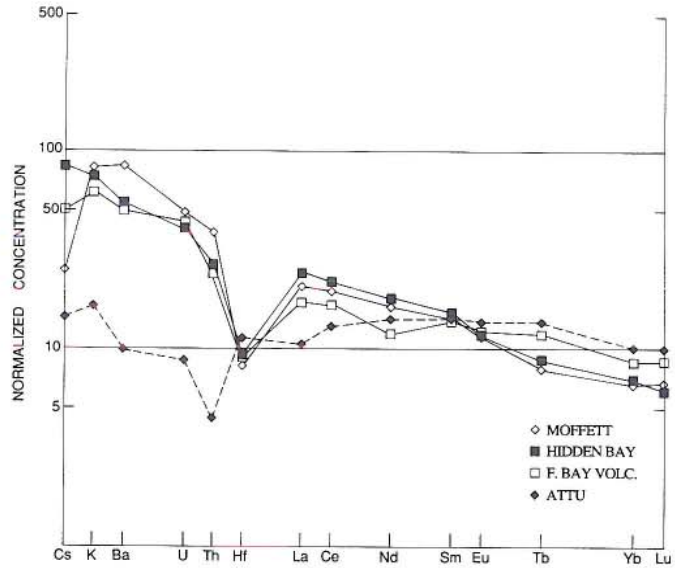


Figure 19. Trace-element patterns of Tertiary to Recent Aleutian mafic magmatic rocks normalized to chondrites and ocean ridge basalts (see Table 1 for normalization values). Trace-element patterns of Adak Island samples, which include a basalt from the Eocene Finger Bay volcanics (CMB-23), a gabbro from the Oligocene Hidden Bay Pluton (HB7-10), and a high-Al basalt from Moffett Volcano (MOF15), show the arc character of all samples studied from the central Aleutians. Note the increase in U, Th, and Ba relative to La in the Moffett lava compared to the older samples. In contrast, the trace-element pattern of a Tertiary pillow lava from Attu Island (SB80-1A) in the western Aleutians shows a fundamentally different pattern, which is related to a spreading center origin (Rubenstone, 1984; R. Kay and others, 1986). Data are given in Tables 1 and 4.

GEOCHEMICAL CHANGES THROUGH TIME IN CENTRAL ALEUTIAN MAGMATIC ROCKS

Temporal changes in the geochemistry of Aleutian magmas indicate processes occurring in the arc crust and mantle. During the existence of the Aleutian arc, the lithosphere (mainly crust) has changed because of the continued intrusion of arc magmas, and the exchange between these arc magmas and the preexisting lithosphere. If disruption of the crust by major folding and thrust faulting is minor (as suggested by field data), the crust that was under a section of arc at its inception is there today, and contains a record of the chemical growth and evolution of the arc. In contrast, the down-going slab and the asthenospheric mantle are continually renewed, and record only a transient situation. With this in mind, we will discuss the magmatic evolution of Adak Island in the middle of the modern Andreanof segment (Fig. 1).

TABLE 5. REPRESENTATIVE ANALYSES OF TERTIARY MAFIC MAGMATIC ROCKS FROM ATTU AND ADAK ISLANDS, ALEUTIAN ISLANDS, ALASKA*

Sample	SB80-1A	CM8-23A	BW8-R36	FB53	FB97	HB7-10
SiO_2	49.70	49.20	48.58	48.49	50.72	51.17
TiO_2	1.46	1.05	0.93	0.91	0.95	0.94
Al_2O_3	15.49	17.90	16.56	19.86	19.10	17.70
Fe_2O_3						
FeO	8.94	9.73	10.40	8.48	9.54	8.42
MnO	0.12	0.21	0.24	0.13	0.22	0.14
MgO	8.52	7.65	6.16	4.86	4.24	6.81
CaO	11.47	10.38	11.73	12.82	9.87	10.42
Na_2O	3.82	2.51	3.60	2.66	3.79	3.16
K_2O	0.23	0.87	0.76	0.81	1.38	1.04
P_2O_5	0.13	0.34	0.29	0.18	0.30	0.20
Total	99.88	99.84	99.25	99.20	100.11	100.00
La	3.90	6.46	9.60	7.66	10.10	8.99
Ce	12.70	16.30	22.70	18.10	23.30	21.28
Nd	10.10	8.66	14.60	15.30	16.50	12.82
Sm	3.22	2.64	3.81	3.01	3.81	3.43
Eu	1.190	1.030	1.140	0.894	1.310	1.023
Tb	0.811	0.692	0.616	0.449	0.563	0.512
Yb	2.47	2.15	1.65	1.30	1.60	1.74
Lu	0.385	0.330	0.260	0.197	0.262	0.239
Rb	5	17	14	18	25	28
Sr	183	432	900	728	753	736
Ba	39	188	148	233	329	214
Cs	0.19	0.67	0.31	0.32	0.73	1.09
U	0.12	0.65	0.49	0.76	0.95	0.61
Th	0.22	1.16	0.98	1.47	1.60	1.37
Hf	2.47	1.99	1.64	1.96	2.59	1.92
Sc	36.2	41.4	43.9	29.9	25.9	30.7
Cr	408	59	255	72	22	225
Ni	185	60	114	36	21	75
Co	408	44	45	40		34
FeO/MgO	1.05	1.27	1.69	1.74	2.25	1.24
Ba/La	10.00	29.10	15.42	30.42	32.57	23.80
La/Sm	1.2	2.4	2.5	2.5	2.7	2.6
La/Yb	1.6	3.0	5.8	5.9	6.3	5.2
Eu/Eu*	0.99	0.99	0.99	0.99	0.99	0.99
Th/Hf	0.09	0.58	0.59	0.75	0.62	0.71
Ba/Hf	16	94	90	119	127	111
Ba/Th	176	162	152	159	206	156
La/Hf	1.6	3.2	5.9	3.9	3.9	4.7
Th/U	1.84	1.80	2.00	1.93	1.68	2.25
$^{87}Sr/^{86}Sr$	0.70357		0.70306	0.70303	0.70326	0.70305
ε_{Nd}	10.10		7.60	9.30	8.40	9.40

*SB80-1A is a pillow lava from the Sarana-Chichagof region, Attu (analyses from Rubenstone, 1984); CM8-23A and BW8-R36 are volcanic clasts from the Finger Bay volcanics, Adak (analyses from Rubenstone, 1984); FB53 and FB97 are gabbros from the Finger Bay pluton, Adak (partial analyses in Kay and others, 1983); HB7-10 is a gabbro from the Hidden Bay pluton, Adak (see Citron and others, 1980). Isotopic analyses are from R. Kay and others (1986).

This discussion is representative of the central Aleutians near Adak Island, not as a model for the entire arc.

Geochemical studies have been made on the Tertiary to Recent magmatic units that crop out on Adak. The units studied in detail include the pre-Oligocene Finger Bay volcanics (Coats, 1956; Rubenstone, 1984) and Finger Bay pluton (Coats, 1956; Kay and others, 1983), the calc-alkaline Oligocene Hidden Bay pluton (Fraser and Snyder, 1959; Citron, 1980; Citron and others, 1980; Kay and others, 1990), and the Pliocene-Recent calc-alkaline volcanoes, Moffett and Adagdak. The distribution of these units is given in Kay and others (1983, 1990), and their geologic setting and history are summarized by Vallier and others (this volume).

The chemistry of the older Adak magmatic rocks is compared to those of the younger Moffett volcano in Table 5 and Figures 19 to 21. The Finger Bay volcanics unit, which shows variable zeolite- to greenschist-facies metamorphism, consists of dominantly mafic to intermediate composition pyroclastic rocks with some flows, dikes, and sills. Analyzed samples include volcanic clasts, flows, and dikes (Rubenstone, 1984). The small (3 km across) tholeiitic Finger Bay pluton is composed mainly of cumulate and noncumulate gabbro with lesser amounts of quartz monzodiorite and other intermediate units (Kay and others, 1983). In contrast, the larger (15 km across), calc-alkaline Hidden Bay pluton is composed dominantly of quartz monzodiorite and granodiorite (58 to 63 percent SiO_2) with lesser amounts of gabbro and leucogranodiorite (Citron, 1980).

In a broad sense, the petrology and geochemistry of the older Adak magmatic rocks are similar to those of the younger Aleutian arc volcanic rocks. The Tertiary calc-alkaline plutons (Fig. 1) on Unalaska (Perfit, 1977; Perfit and others, 1980a) and Kagalaska Islands (Citron, 1980; Kay and others, 1982) to the east, and Amchitka Island (Kay, 1980) to the west are also geochemically similar to the Adak rocks, particularly to the rocks of the Hidden Bay pluton (Citron, 1980; Citron and others, 1980; Kay and others, 1982; Kay and others, 1990). Like the younger arc volcanic rocks, none of these older magmatic rocks is chemically similar to the island-arc tholeiites of the western Pacific (Kay and others, 1982, 1983; Rubenstone, 1984).

The early mid-Tertiary magmatic rocks in the central and eastern arc are distinct from the early Tertiary volcanic and plutonic rocks on Attu Island in the volcanically interactive western Aleutian arc (Rubenstone, 1984; Shelton, 1986; R. Kay and others, 1986). These old Attu magmatic rocks appear to come from a spreading ridge environment and are geochemically like ocean-ridge rocks, as shown by their flat or light REE depleted REE patterns, lack of alkali and alkaline earth enrichment, lack of HFSE depletion relative to the REE (Fig. 19), and MORB-like ϵ_{Nd} (Table 5). Younger Tertiary volcanic rocks on Attu Island are more arc-like in character (Rubenstone, 1984; R. Kay and others, 1986).

Specific differences in the geochemistry and mineralogy between the Adak andesitic to dacitic composition volcanic and plutonic rocks can be explained by differences in the conditions of their crystallization. The plutons appear to represent the last unerupted stages of the volcanoes as suggested by mineral assemblages, which indicate more water-rich, cooler crystallization conditions (low-Al amphibole, biotite, quartz, and potassium feldspar; S. Kay and others, 1986b, 1990). Some plutonic rocks have lower Ba/La ratios at higher La/Yb ratios than observed in any of the young Aleutian volcanic rocks (Fig. 20), a trend that is consistent with an increased role for amphibole fractionation in the plutons, leading to higher La/Yb ratios and lower Ba/La ratios in the plutonic rocks.

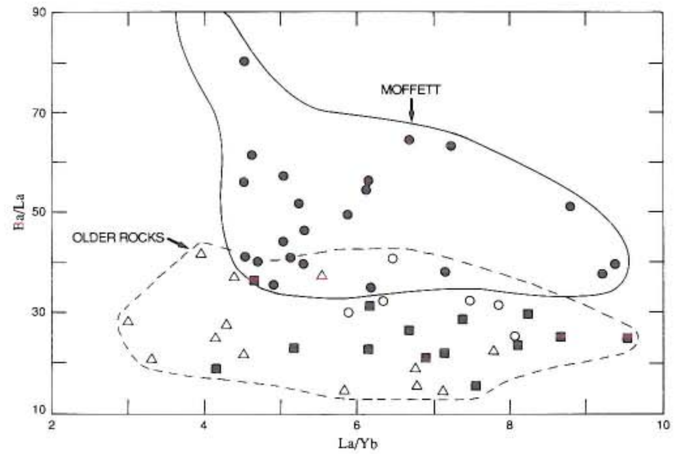


Figure 20. Plot of Ba/La versus La/Yb for Adak Island magmatic rocks from the Eocene Finger Bay volcanics (triangles), the Eocene(?) Finger Bay pluton (open circles), the Oligocene Hidden Bay pluton (solid squares), and the Plio-Pleistocene Moffett Volcano (solid circles). High-Mg andesites of Kay (1978) plot off-scale at Ba/La ~10 and La/Yb ~50. Although La/Yb ratios of the Moffett volcanics are similar to those of the older rocks, the Ba/La ratios of the Moffett volcanics are distinctly higher, suggesting a change in the source region beneath Adak Island over this time period. Except for the Finger Bay volcanic samples that have had La/Yb ratios >6 combined with Ba/La ratios <20, the older samples have Ba/La and La/Yb ratios that overlap those of Plio-Pleistocene to Recent volcanic rocks in other parts of the arc (compare with Figs. 15A and B).

In contrast, differences in incompatible and only slightly compatible element ratios (e.g., Ba, the light REE) in the Adak rocks record the evolution of magmas added to the crust on Adak. Some of these differences are illustrated in Figure 20, a plot of Ba/La (incompatible elements) versus La/Yb ratio (where Yb is more compatible than La). For example, the Ba/La ratios of most of the Tertiary samples are lower than those of samples from Moffett volcano, but are within the range of other recent Aleutian arc volcanic rocks (see Fig. 15B and recall that Moffett samples have relatively high Ba/La ratios). Even if amphibole fractionation has altered the Ba/La and La/Yb ratios in some of the plutonic rocks, the magnitude of the Ba/La ratios in the plutonic rocks as a whole still suggests that their source region, like that of the Finger Bay volcanic rocks, had a lower Ba/La ratio than the source region of the Moffett volcanic rocks.

The oldest Adak samples, those from the basaltic Finger Bay

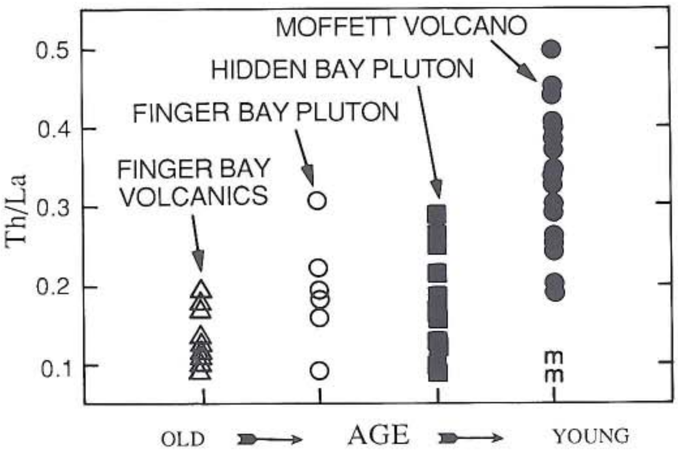


Figure 21. Plot of Th/La ratios versus relative age for Adak magmatic rocks. In addition to the symbols used in Figure 18, m represents the high-Mg andesites. The Th/La plot is representative of the high ratios of Th, U, Ba, Sr, and Rb to the other incompatible elements in many of the younger Moffett rocks relative to the older units.

volcanics unit, fall in two general groups as shown in Figure 20 (data from Rubenstone, 1984). The first group has Ba/La and La/Yb ratios that plot outside the Moffett field, but within the modern arc field. The isotopic characteristics of the one analyzed sample (ϵ_{Nd} = 6.8 and $^{87}Sr/^{86}Sr$ = 0.70332) from this group are like modern arc lavas (R. Kay and others, 1986). The second group has lower Ba/La (most are 15 to 20) at higher La/Yb ratios (~6 to 8) than observed in modern Aleutian basalts. Although the isotopic characteristics of this second group (3 samples, ϵ_{Nd} = 7.6 to 7.7 and $^{87}Sr/^{86}Sr$ = 0.70296 to 0.70316, R. Kay and others, 1986) are also within the range of modern Aleutian arc lavas, they are more oceanic-like than the first group. The subtle, somewhat more OIB-like character (lower Ba/La ratios combined with higher La/Yb ratios) of some of the second group of two samples could be consistent with their eruption in the early stages of the arc, especially if the initiation of the modern arc is related to an old fracture zone (Vallier and others, this volume). However, the similar characteristics of the Finger Bay volcanics and the modern arc lavas is much more striking than their differences (Rubenstone, 1984).

Comparison of other incompatible trace-element ratios between the older Adak rocks and those from Moffett Volcano shows that the older rocks have a lower U, Th, and Ba content relative to the other incompatible elements. Thus, ratios such as Th/La (see Fig. 21), Th/Hf, Ba/La, and Ba/Hf are generally lower in the older rocks than in the Moffett lavas, whereas ratios like Ba/Th show no obvious age correlations. The most notable exceptions are a few Finger Bay volcanic samples that have high Ba/Th ratios. These samples also have the highest Ba/La and Ba/Hf ratios, suggesting that they have higher Ba than the other Finger Bay volcanic samples. Isotopic data are not available on any of these samples.

Examination of the available isotopic data on Adak and Kagalaska plutonic rocks shows that ϵ_{Nd} in the plutonic samples (ϵ_{Nd} = +7.1 to +9.3) is, in some cases, higher than in any of the modern arc volcanic rocks (ϵ_{Nd} = +7.3 ± 1.25 for 56 samples; R. Kay and others, 1986). For example, samples of the Hidden Bay gabbro and the Finger Bay plutons have ϵ_{Nd} ranging from 8.4 to 9.4, and R. Kay and others (1986) suggest that these high ϵ_{Nd} values may result from assimilation of old oceanic crust by the plutonic magmas in the basement of the Adak crust. However, the process must be more complex, because ϵ_{Nd} decreases (9.4 to 7.1) and $^{87}Sr/^{86}Sr$ increases (0.70305 to 0.70328) with increasing SiO_2 in the three samples from the Hidden Bay pluton that have been analyzed. Although a systematic trend with SiO_2 content is observed in the Hidden Bay pluton data, no pattern emerges from the plutonic rock data as a whole (Kay and others, 1990).

As observed regionally for the young volcanic rocks, the higher ϵ_{Nd} in the older rocks appears to correlate in a general way with lower Ba/La, Ba/Hf, and Th/Hf ratios, suggesting that a component with relatively high Ba, Th, and U, and low ϵ_{Nd}, contributes to the distinctive trace-element signature of the younger rocks of the Adak region (e.g., Moffett Volcano). Contamination of the Moffett lavas with older Adak crustal rocks cannot easily account for the characteristics of this component, suggesting that the change is the result of change or addition of a component in the subcrustal source. The data on the older Adak rocks suggest that this component (perhaps a distinctive sediment type) became important after the mid-Tertiary and could be related to the formation of a well-defined accretionary prism in this part of the arc at about 5 Ma (Scholl and others, 1987).

ALEUTIAN MAGMATIC PROCESSES: AN INTEGRATED VIEW

The preceding sections have emphasized that magmatic and tectonic processes can be inferred from the mineralogy and the chemical and isotopic compositions of Aleutian igneous rocks. We have integrated these processes on a regional scale, to derive a tectonic and lithologic cross section of the Aleutian arc, as shown in Figure 22. This section is drawn across the middle of a coherent arc segment, such as near Adak Island. The discussion proceeds from deep to shallow tectonic levels.

Mantle level

We call our hypothesis for the formation of Aleutian arc magmas the convective limb process. In this process, a melt or solute-rich fluid (Eggler, 1989), which includes components from sediment and hydrothermally altered oceanic crust, is transferred from the subducted plate into the overlying mantle peridotite, causing the mantle to partially melt and rise as a broad convective limb (Fig. 22). The melt percolates continuously through the rising peridotite over a depth range from 100 to perhaps 30 km, where it segregates and ponds under the lithosphere, which is largely crust. A porous flow mechanism (Turcotte, 1987) appears to operate in favor of hydrofracture, because Aleutian mantle xenoliths have textures indicating penetrative deformation. Diapirism also appears to be ruled out, because the penetration velocity of hypothetical diapirs is slower than the segregation velocity of the interstitial melt required for buoyancy. Broadly speaking, the convective-limb hypothesis satisfies the observational demands of the distribution of seismic and velocity minima in the mantle over the plate (Engdahl and Billington, 1986, and Engdahl, oral communication, 1987) and the compositional coincidence of the least evolved arc magmas (olivine tholeiites) and partial melts of the spinel lherzolite mantle at the 30-km depth (e.g., Fujii and Scarfe, 1985).

Evidence that subducted sediment and hydrothermally altered oceanic crust is involved in the source region of Aleutian magmas comes from trace elements, radiogenic isotopes, and ^{10}Be. However, oxygen isotopic values of Aleutian arc lavas are close to those of MORB, indicating that the bulk of the Aleutian magmas comes from the mantle. Regional variations in incompatible trace-element and isotopic abundances in Aleutian arc volcanoes, as well as some of the heterogeneity within single centers, probably results from variability in the subducted component. This interpretation, however, requires a more extensive evaluation of the regional differences in the crust of the Pacific Plate and of the arc basement.

In major-element composition (except K_2O), the most mafic Aleutian basalts are similar to MORB, and the proposed convective-limb is similar to the process for magma generation and migration thought to occur at the mid-oceanic ridges (e.g., Turcotte, 1987). However, the driving force in the arc—H_2O-CO_2 flux-created buoyancy—is distinctive. The observation that the distance between volcanoes both along arcs and at mid-oceanic ridges is about the same supports our speculative model that these characteristic volcanic spacing result from transverse convective rolls. In this model, following the release of the basaltic fraction from the upwelling convecting limb, residual mantle largely descends both away from the trench and between the volcanic centers (not shown in Fig. 22). A quite different model commonly cited (e.g., Marsh, 1979) is that the spacing of arc volcanoes is due to Rayleigh-Taylor instabilities within a magma ribbon over the subducted plate at 100 km.

An important difference between arc and MORB tectonic regimes is that the magma-production rate (mass of magma per length of arc or ridge per time) is generally lower in arcs than at mid-oceanic ridges. However, the magma production rates for the least productive mid-oceanic ridges and the most productive arcs approach each other, suggesting a continuum of pressure-release melting mechanisms. For example, an Aleutian magma-production rate averaged over 60 m.y. is 35 $km^3/km/m.y.$ (calculated from the crustal section of Grow, 1973, as modified by S. Kay and Kay, 1985b), which compares closely with magma production in the slowest mid-ocean ridges (southwest Indian Ridge: 40 $km^3/km/m.y.$, calculated from an 8 mm/yr spreading rate and a 5-km-thick crust; Fisher and Sclater, 1983).

The role of volatiles released from the subducted plate is crucial in the production-rate differences of arc and MORB lavas. The higher volatile content of arc magmas means that the same percentage of melting is achieved at lower temperature in arcs

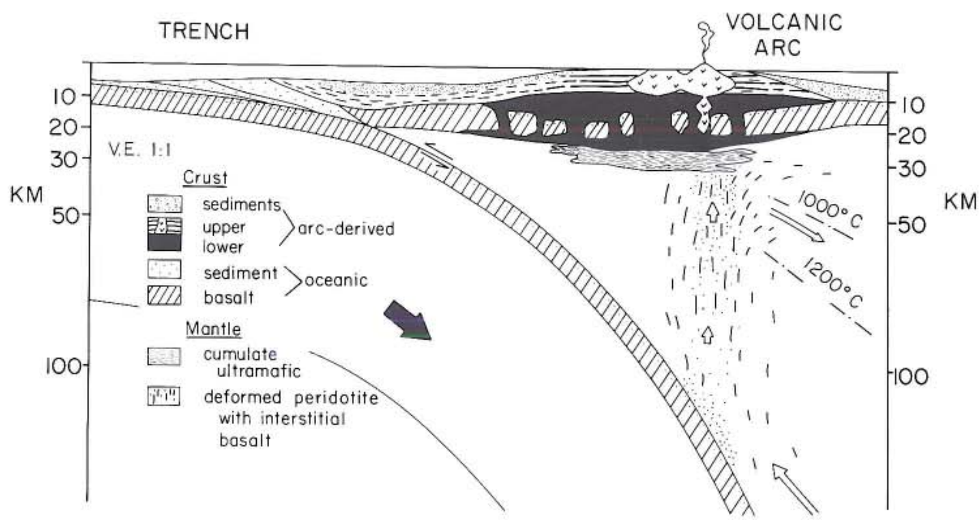


Figure 22. Cross section (no vertical exaggeration) showing crustal formation in the central Aleutian arc (modified from S. Kay and Kay, 1985b; R. Kay and others, 1986). Configuration of subducted plate, crustal thickness, and forearc lithologies are constrained by geophysical and seismological data (Grow, 1973; Engdahl and Billington, 1986). Melting (represented by dots) occurs in peridotite that lies above the subducting plate. Convective rise of the buoyant peridotite causes furhter melting (by decompression) accompanied by percolation of melt through the denser peridotite. Segregated mafic minerals from ponded olivine tholeiite form the cumulative ultramafic rocks that occur at the crust-mantle boundary. A basic lower crust consists of unassimilated residues of crystalline oceanic crust and intrusive hi-Al basalt and its mafic crystalline fractionates. Intermediate composition upper crust consists of igneous intrusions and extrusions of mafic to intermediate composition.

than at mid-oceanic ridges. For arcs, in the extreme cases of no released volatiles or of temperatures that are too low in the hanging wall, no melting of overlying peridotite occurs. In cases where the peridotite is hot enough to melt, volatile release drives buoyancy. If the magma-production rate does not vary directly with the release rate of volatiles into the mantle at the depth required for melting (100 km), then a variable volatile content of magmas could result (as suggested above for the Aleutian tholeiitic versus calc-alkaline magmas). Volatiles may also be stored in the hanging wall over the subducted plate.

Magmatic segregation processes in the mantle under the Aleutian arc have an important bearing on mantle evolution. Assuming that some continental crustal components in sediment and altered oceanic crust are recycled from the mantle back to the crust as components of arc magmas, any inefficiency of the recycling process will leave crustal components in the mantle. This inefficiency probably occurs at three sites: (1) during extraction of crustal components from the subducted plate, (2) during extraction of primitive basaltic melt from peridotite, and (3) during fractionation of basalt in the mantle or more likely at the crust-mantle boundary. Returned to the mantle at all three sites are mafic or ultramafic residues or fractionated liquidus minerals that are at equilibrium with arc magmas and therefore share their isotopic signatures. Some trace-element ratios in these arc-related mantle rocks are dominated by the recycled component. The production rate of these residues is large—perhaps one-fifth of the production rate of MORB residues, which is about four times the ratio of magma production rates reflecting lower percentages of melting at arcs (Kay, 1980). The return mechanism of the residues to deeper levels in the mantle is obvious for the subducted plate, but less apparent for the mantle residues and for the cumulates that pond immediately below the crust-mantle boundary. A delamination mechanism may apply to the cumulates.

Crustal level

The crust-mantle boundary is considered to be magmatic, and coincides with a series of short-lived magma ponds whose bases are formed by the accumulation of olivine and clinopyroxene—the liquidus phases of olivine tholeiite. Gabbros, together with crystallized high-Al basalts and their accumulated liquidus phases, constitute the lowermost crust. Because the present asthenospheric mantle reaches the crust-mantle boundary, oceanic mantle lithosphere that was under the arc at its inception has flowed away from under the volcanic line. Similarly, ultramafic cumulates from earlier stages of arc magmatism have also flowed away.

If a continuous elongated pool of magma accumulates at the crust/mantle boundary under the volcanic line, Rayleigh-Taylor instability could result in volcanic spacing. However, the magmatic production rate is probably too low to create a long-lived pool at such a shallow depth, because it would crystallize—a process that could be accelerated by crustal assimilation. Also, chemical and isotopic heterogeneity at the highest volume centers such as Okmok indicates incomplete mixing of magma batches—an observation that is difficult to reconcile with a magma chamber extending along the arc at the crust-mantle boundary.

Within the present mid- to lower crust beneath the arc, some of the original oceanic crustal basement probably remains (S. Kay and Kay, 1985b). This oceanic crust is most likely to be preserved in the large (about 500 km) coherent crustal segments. Pieces of this oceanic crust, metasomatized by arc intrusive bodies, occur as xenoliths in Tertiary dikes on Kanaga Island (Swanson and others, 1987). The hydrated upper layers of this oceanic crust, as well as sediments on top of it, should undergo dehydration and melting in response to the heating that accompanies magmatic intrusion, and can be recycled upward within the arc crust. Arc volcanic flows associated with the early stages of arc formation may also overlie the oceanic crust in the deeper parts of the arc crust.

Within coherent arc segments (the region shown in Fig. 22), the mid- to lower crust is the site of magma chambers in which the mixing and fractionation processes occur that yield the calc-alkaline series of rocks (Fig. 9). These processes imply that an important component of the crust at this level is the crystal residue from the fractionation processes that produpce the erupted calc-alkaline lavas. These magma chambers probably did not exist until the arc crust was built to some minimum thickness in the Tertiary (Kay and others, 1990). The upper crust in these regions consists of the plutons, which represent the last stages of calc-alkaline volcanoes, and the erupted volcanic rocks.

In the regions between coherent crustal blocks, the large volcanoes that are not in linear belts are attributed to extensional stress regimes (Fig. 9). If crustal dilation occurs in these regions, the preexisting oceanic crust may not be present, and either the crust is thinner or there is a higher intrusion rate. In either case, the arc crust probably consists of a large proportion of arc-derived magmas. Differentiation occurs along a tholeiitic trend, because fractionation and any mixing that occurs are generally happening at high temperatures at shallow levels. Remelted, hydrothermally altered roof rocks may be important in the genesis of some members of the tholeiitic series. Melting at shallow levels is facilitated by the high-temperature Mg-rich magmas, which can reach the surface in these regions.

Growth of the continental crust

As implied in this paper, the mean composition of the Aleutian arc crust is high-Al basalt (or more mafic basalt, depending on the proportion of preexisting oceanic crust). In contrast, the bulk composition of the continental crust is usually considered to be andesitic (e.g., Taylor and McLennan, 1985), similar to the composition of only the upper crust of the Aleutian arc. This creates a compositional dilemma if continental crust is composed of accreted arcs. As a solution to this problem, we have proposed (S. Kay and Kay, 1985b; and R. Kay and Kay, 1986, 1988) that lower crustal delamination accompanies structural suturing of oceanic arc terranes such as the Aleutians to existing continental margins, and that the mafic lower crust of arcs is recycled back into the mantle.

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NOTES ADDED IN PROOF

Our chapter, written in 1987, and slightly updated in 1989, was organized around three current controversies. What are the parental magmas in the Aleutians? What accounts for the trace element and isotope characteristics of the magmas? What are the underlying causes of petrologic and geochemical diversity of the magmas? The topics we covered did not include some found in Fournelle and others (this volume), which was written in 1991. The most important information in this update is a list of additional references. As the three controversies of 1987 continue at the time of this update (9/93), we will summarize our current views, which often contrast with those in Fournelle and others. We have not read their update.

Aleutian parental magmas

We recognized two types of parental (mantle-derived) magma types for which there was tangible evidence: a common basalt and an Mg-andesite. Alternatives for the origin of the basalt were: (1) melting of subducted mafic oceanic crust (with eclogite mineralogy) yielding high-Al basalt or (2) melting of peridotite of the mantle wedge, yielding high-Mg basalt. Our preference for the second alternative has been strengthened by recent experimental work (e.g., Rapp, 1990; Draper and Johnston, 1992). A critical result from the new experimental studies is the confirmation that at lower crustal conditions, high-Al basalt can result from fractionation of high-Mg basalt. For the overwhelming majority of experimental petrologists, arc basalt genesis by peridotite melting is now the choice, as it was for us in 1987.

The mechanics of the melting has been a topic of great interest, with a polybaric extraction of melt from a rising peridotite mass ("melting column") being popular (e.g., Plank and Langmuir, 1988). Our Figure 22 shows the "melting column." The along-arc variations within the mantle in a cross section like Figure 22 were described by us as "transverse rolls." Even given the popularity of models like that in Figure 22 (and alternatives), the problem of the thermal and flow structure of the subarc mantle remains unresolved. Progress is impeded by doctrinaire positions like that of Fournelle and others, who dismiss our model with the observation that our "mantle flow is opposite that of all geophysical models. . . . " This is not true, as evidenced by a careful reading of Davies and Stevenson (1992) and earlier references cited by them. Ultimately, tomographic work like that of Hasegawa and others (1991) in Japan will be needed to constrain any spatial patterns.

Despite its rarity, the Mg-andesite first reported in Kay (1978) has enjoyed a resurgent popularity due to the efforts of Defant and Drummond (1990). These authors coined the name "adakite" for the locality on Mt. Moffett, Adak Island, and concurred with our interpretation that it is a melt of the subducted oceanic crust. This remarkable magma, with the Nd, Pb, and Sr isotope ratios of MORB, is quite distinct from the magmas of the main Aleutian volcanoes, a fact not appreciated by Fournelle and others.

Finally, over the last 3 years we have investigated a third magma type, with chemical similarities toward boninites (with higher SiO_2 at the same Mg number compared to the Mg-basalt). We have concentrated on the locality at Piip volcano (Seliverstov and others, 1990; Volynets and others, 1993a; Yogodzinski and others, 1994). Piip is an active submarine volcano situated within a graben immediately behind the Komandorsky Islands, the westernmost islands of the Aleutian chain. The unique chemistry of the primary magma is attributed to shallow melting of hydrous peridotite. Recycled crustal components are much less abundant in these magmas than those of the Aleutian arc to the east. A subducted oceanic plate moving nearly parallel to the arc appears to lie under Piip (Boyd and Creager, 1991).

Trace element and isotopic characteristics

The nature of various components and their proportions in the mantle wedge is the most popular topic in arc magma genesis. In the Aleutians, a three-component model (slab igneous rock, sediment, depleted mantle) continues to be popular (e.g., Miller and others, 1992, which we look upon as an update to Kay, 1980, although this reference is not cited). The parallel between these two papers (which both contained models of Okmok basalt) extends to both using, as a component added to the mantle wedge, a melt of the basaltic part of the subducted slab. The weight of present-day opinion (except for the "adakite" mentioned above) comes down against this slab-melt component, in favor of a fluid component.

Mass balance calculations of ^{10}Be in arc magmas have spurred continued quantitative evolution of element inputs and outputs (recycling) in arcs, through correlations with elements like Ba (see our Fig. 18) and B (Morris and others, 1990). Sediment analyses relevant to the recycling problem (e.g., oceanic sediments south of the Aleutian trench) have been reported by Plank and Langmuir (1993) following the extensive analyses of Kay and Kay (1988).

Finally, the discovery of a geochemically distinct region in the westernmost Aleutian and Komandorski Islands (Yogodzinski and others, 1993) has caused a reassessment of common assumptions regarding the MORB-like and depleted mantle components. The extremely nonradiogenic Pb isotope ratios in "adakite" and boninitic magmas of the region strongly imply that mantle as well as crustal components vary along the arc.

Magmatic diversity

We recognized Aleutian magma series of two end member types: calcalkaline (CA) and tholeiitic (TH). We also recognized the full range of transitional types. We put great emphasis on the crustal environment as controlling the development of these magma series, as ultimately we related all magmas to mantle-derived parental olivine tholeiite. This was even true with CA plutons (Kay and others, 1990) although we held that the "relation" involved melting, assimilation, mixing, and fractional crystallization processes. Singer and others (1992) exploit the idea of extensional crustal environment (due to block rotations, see Geist and others, 1988) in their paper on the TH Seguam volcanic center. Romick and others (1992) show by trace element modelling of central Aleutian CA ashes that fractionation of a mid-crustal (amphibole-bearing) mineral assemblage was detectable even for magmas with upper crustal (amphibole-free) phenocryst populations. Romick and others (1992) found that some Aleutian dacites and their phenocrysts are quite similar to amphibole-bearing Mt. St. Helen's magmas. Not to see amphibole in Aleutian magmas is to practice selective vision.

Miller and others (1992) evaluate alternatives to tectonic controls of Aleutian magma series. They review the application of melt-peridotite interaction (e.g., Kelemen, 1990) in producing distinct olivine tholeiite parental magmas. They favor the view that "differences between CA and TH volcanoes originate in the slab, rather than being controlled by the stress regime in the crust." Specifically, the TH parent magmas are higher percentage partial melts associated with volatile-rich subducted oceanic crust of fracture zones. This contrasts with our view of the along-arc geochemical variability (see our Fig. 15) in which both CA and TH centers from a region have common parental magmas. The geochemical contrasts between the Islands of Four Mountains segment and the Andreanof segment are particularly obvious. Both these segments contain both CA and TH centers.

Finally, new data on Aleutian back-arc volcanism is reported by Volynets and others (1993b) for volcanic rocks dredged from seamounts north of the eastern Aleutian Islands. A cross-arc traverse that includes low-K magmas of the Islands of Four Mountains (see our Appendix on microfiche [back of the volume] and Singer and others, 1992) and high-K magmas erupted over deeper parts of the same subducted slab has been accomplished.

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