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Chapter 23

Age, character, and significance of Aleutian arc volcanism

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INTRODUCTION

The Aleutian volcanic arc stretches nearly 3,000 km, from the Commander Islands off Kamchatka (U.S.S.R.), along the southern Bering Sea margin, and across the continental margin onto the Alaskan landmass. Intimately associated with the arc is the subduction of the Pacific Plate beneath the North American Plate. To the east, subduction is nearly orthogonal but becomes increasingly oblique westward. Near Buldir Island, motion between the two plates becomes strike-slip and volcanism ceases. The Aleutian volcanic front, which has in many places remained nearly fixed for at least several tens of millions of years, contains about 80 major volcanic vents, half of which have been historically active. These vents have yielded a spectrum of rock types, from basalt through andesite to dacite and rhyolite. This diversity of rock types is present throughout the history of nearly all the volcanic centers, almost regardless of volcano size and age.

Several features of the Aleutian volcanic arc—a long history of fixed volcanism extending from continental to oceanic crust, the focusing of large amounts of thermal energy on small areas of crust for long periods of time, the pattern of changing convergence, as well as the diversity of rock types—present an excellent opportunity to study the connection between global tectonics, magmatism, and continent evolution. The study of Aleutian volcanism can shed light both on deep-seated magmatic processes and on the interplay of the chemistry and physics of magma evolution, and particularly on the near-surface behavior of magma in various local tectonic and thermal regimes.

This chapter focuses primarily on the latest episode (the last 1 to 2 m.y.) of subaerial volcanism of the oceanic segment of the Aleutian arc, that is, the western and central parts of the volcanic arc, with some discussion of relevant features of volcanism on the Alaska Peninsula. Locations, physical characteristics, and historic eruptions of these volcanoes are given in Table 1.

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HISTORY OF GEOLOGICAL EXPLORATION

The harsh climate of the Aleutian Islands and the treacherous nature of the surrounding sea have made exploration there difficult, even today. In the 20 years following the discovery voyage (1741) by Bering and Chirikof, promyshleniki (frontiersmen) found every Aleutian island in their relentless harvest of sea otters and foxes. Beginning in 1760 with the Russian occupation and the appearance of missionaries, there were written accounts of volcanic eruptions (e.g., Veniaminof, 1840). Black (1981) chronicles some Aleut oral accounts of volcanic activity.

T. Jaggar visited Bogoslof Island one year after it erupted in 1906 (Jagger, 1908). He also visited Atka, Umnak, Unalaska, and Akutan Islands, as well as Pavlof volcano on the Alaska Peninsula. In the 1930s, R. H. Finch of the Hawaiian Volcano Observatory explored some of the eastern volcanoes (1934, Shishaldin; 1935, Akutan). At about the same time, Bernard Hubbard, the so-called "Glacier Priest", ascended Shishaldin and Veniaminof Volcanoes and explored Katmai and the giant Aniakchak Caldera (Hubbard, 1932a, 1932b, 1936).

The most comprehensive geologic field program in the Aleutians was that of the U.S. Geological Survey between 1945 and 1954. From this project, geologic reports were published on 13 of the central and western Aleutian Islands, and on two volcanoes of the Alaska Peninsula. These reconnaissance studies were published as *U.S. Geological Survey Bulletin 1028.* Much of our knowledge of Aleutian geology stems from that project. The most comprehensive study of any single volcanic event in the entire arc has been of the 1912 Katmai eruption (e.g., Griggs, 1922; Fenner, 1923, 1926; Curtis, 1968; Hildreth, 1983, 1987; an issue of Geophysical Research Letters, August 1991; Miller and Richter, this volume).

With the recognition of the relation between subduction and volcanism at convergent margins, the Aleutian island arc became the focus of renewed interest in the 1970's. Since then, six previously unmapped volcanic centers have been investigated. These centers include Amak Island (Marsh and Leitz, 1979), northern Akutan Island (Romick, 1982; Romick and others, 1990), northern Atka Island (Marsh, 1980), Seguam Island (Myers and Sing-

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TABLE 1: LOCATIONS, PHYSICAL CHARACTERISTICS, AND DATES OF HISTORIC ERUPTIONS OF ALEUTIAN VOLCANOES WEST OF 162°W*

Island/Center	Position	Elevation (m)	Dimensions (km)	Historic Activity
Amak Island	163.15W, 55.42N	513	5.0 x 4.2	1700-10
Cold Bay Volc. Complex			31.3 x 23.4	
Frosty Peak	162.77W, 55.08N	1920	17.4	Holocene
Mt. Simeon	162.75W, 55.19N	287	3.0 x 2.0	
N. Walrus Peak	162.81W, 54.98N	893		
S. Walrus Peak	162.18W, 54.98N	892	15.9 x 10.9	
Unimak Island			110.7 x 56.5	
Roundtop	163.60W, 54.80N	1871	7.5	Holocene
Isanotski Peaks	163.73W, 54.75N	2446	11.2	1795, 1825, 1830, 1831, 1845
Shishaldin Volc.	163.97W, 54.75N	2857	17.4	1775-78,1790,1819,1824,1825,1826,1827-29,1830-31,
				1838,1842,1865,1880-81,1883,1897,1898,1899,1901,
				1912,1922,1925,1928,1929,1929-32,1932,1946,1946-47,
				1948,1951,1953,1955,1963,1967,1975,1976,1978, 1979
Fisher Caldera			17.4 x 11.2	1825,1826-27,1830
Mt. Finch	164.36W, 54.67N	478	2.8	
Eickelberg Peak	164.46W, 54.68N	1114	2.8	
Westdahl-Pogromni Volc. Complex			22.5 x 18.8	
Westdahl Peak	164.62W, 54.52N	1560		1964, 1978, 1979
Faris Peak	164.66W, 54.52N	1654		
Pogromni	164.67W, 54.57N	2002		1795, 1796, 1820, 1827-30
Akun Island			21.4 x 17.7	
Mt. Gilbert	165.65W, 54.25N	818		ca. 1834 (fumarole)
Akutan Island	166.00W, 54.13N	1303	29.9 x 21.6	1790,1828,1838,1845,1848,1852,1865,1867, 1883,
				1887,1892,1896,1907,1908,1911,1912,1928,1929,
				1931,1946-47,1948,1951,1953,1972-73,
				1974,1976-77, 1978,1980
Unalaska Island			128.9 x 53.5	
Makushin Volc.	166.93W, 53.90N	2036	38.6 x 17.4	1768-69,1790-92,1802,1818,1826-38,1844,1865,
				1867, 1883,1907,1912,1926,1938,1951,1852, 1980
Wide Bay Cone	166.60W, 53.97N	640		
Table Top Mtn.	166.66W, 53.97N	800		
Sugarloaf	166.73W, 53.94N	611		
Pakushin Cone	166.95W, 53.83N	1035		
Bogoslof Island	169.03W, 53.93N	46	1.5 x 0.5	1796,1804,1814,1820,1884,1890,1891,1906,1907,1909,
				1910.1913.1926. 1917.1931.1951
Umnak Island			118.7 x 27.9	
Okmok Caldera	168.13W, 53.42N	1072	44.8 x 27.4	1805,1817-20,1824-29,1830,1878,1899,1931,1936,
				1938, 1943, 1945, 1983; ca 1986 (Cone A)
Mt. Idak		585		
Jag Peak		900		
Mt. Tulik	168.05W, 53.37N	877		
Mt. Recheshnoi	168.55W, 63.15N	1984	10.0	Holocene
Mt. Vsevidof	168.68W, 53.13N	2076	12.4	1784, 1790, 1817, 1830, 1878, 1880, 1957
Kagamil Island	169.72W, 52.98N	893	10.2 x 6.2	1929
Uliaga Island	169.77W, 53.07N	888	3.7	Holocene
Chuginadak Island			23.4 x 13.9	
eastern	169.75W, 52.80N	1170	14.2 x 13.9	
Mt. Cleveland	169.95W, 52.82N	1730	8.7 x 8.2	1893,1897,1929,1932,1938,1944, 1951,1975, 1987
Carlisle Island	170.05W, 52.90N	1620	8.5 x 7.2	1774, 1828, 1838, 1987
Herbert Island	170.12W, 52.75N	1290	9.2 x 8.5	Holocene

TABLE 1: LOCATIONS, PHYSICAL CHARACTERISTICS, AND DATES OF HISTORIC ERUPTIONS OF ALEUTIAN VOLCANOES WEST OF 162°W* (continued)

Island/Center	Position	Elevation (m)	Dimensions (km)	Historic Activity
Yunaska Island			22.9 x 12.2	
northeast	170.70W, 52.63N	457	12.7 x 11.9	1817, 1824, 1825, 1830, 1873, 1920s?,1937
southwest	170.76W, 52.56N	915	11.9 x 7.0	
Chagulak Island	171.13W, 52.57N	1141	3.2 x 3.2	Holocene
Amukta Island	171.25W, 52.50N	1064	9.5 x 8.0	1786-91, 1876, 1963, 1987
Seguam Island				1786-90, 1827, 1891, 1892, 1902, 1927
· Pyre Peak	172.52W, 52.31N	1054		1977
eastern caldera	172.37W, 52.33N	847		
eastern volcano	172.33W, 52.35N	587		
western center	172.59W, 52.27N	713		
seafloor, N of Amlia	173.50W, 52.00N			1966-67 (submarine: hydrophonic)
Atka Island			98.3 x 34.6	
Korovin	174.15W, 52.38N	1533		1829, 1830, 1844, 1907, 1951, 1977
Klicheuf	174.14W, 52.33N	1451		
Sarichef	174.03W, 52.38N	1056		1812
Konia	174.13W, 52.36N	1125		
Koniuji Island	175.13W, 52.22N	266	1.2	
Kasatochi Island	175.50W, 52.18N	314	2.7 x 2.2	1760, 1827, 1828, 1899?
Great Sitkin Island	176.13W, 52.07N	1740	17.7 x 16.7	1760,1792,1829,1904,1933,1945,1949-50,1950, 1953, 1974
Adak Island			52.2 x 39.1	
Mt. Moffett	176.86W, 51.99N	1196	10.0 x 10.0	
Mt. Adagdak	176.60W, 51.94N	621	4.0 x 3.5	
Kanaga Island			49.8 x 12.4	
Kanaga Volcano	177.15W, 51.93N	1307	10.4 x 6.2	1768,1783-87,1790,1791,1827,1829,1904,1906, 1933, 1942
Bobrof Island	177.43W, 51.90N	738	3.7 x 2.2	
Tanaga Island			42.0 x 37.8	
Gash Bay	177.95W, 51.89N	650		
Takawagha	178.02W, 51.87N	1448	13.7 x 8.7	Holocene
Tanaga	178.15W, 51.88N	1777	5.8 x 5.0	1763-70, 1791, 1829, 1914
Sajaka	178.20W, 51.78N	1304	2.7	
Gareloi Island	178.80W, 51.80N	1573	10.7 x 9.2	1760,1790,1791,1792,1828-29,1873,1922,1929, 1930, 1950-51, 1952, 1980
Semisopochnoi Is.			20.6 x 17.4	
Ragged Top	179.67E, 51.82N	904	3.2	
Anvil Peak	179.60E, 51.99N	1223	3.7	
Mt. Cerberus	179.59E, 51.93N	774	3.7	1772, 1790, 1792, 1828, 1830, 1873, 1922, 1929, 1930
Sugarloaf	179.63E, 51.90N	856	3.5	
Perret Ridge	179.66E, 51.97N	889	5.0	
Little Sitkin Island	178.53E, 51.95N	1202	10.7 x 10.0	1776, 1828
Davidof Island	178.33E, 51.97N	1159	3.5 x 1.2	Holocene
Khvostof Island	178.29E, 51.98N	260	2.7 x 1.5	
Segula Island	178.13E, 52.02N	1140	7.7 x 7.0	Holocene
Kiska Island			41.0 x 15.7	
Kiska Volcano	177.60E, 51.10N	1220	9.0 x 8.2	1907, 1927, 1962-64, 1969
Buldir Island	175.92E, 52.35N	656	7.2 x 4.2	Holocene

*Dimensions are of either island, volcanic center/complex, or volcano; single dimension is diameter, otherwise maximum x minimum dimension.

er, 1987; Singer and others, 1991), Tanaga (Coats and Marsh, 1984), and Shishaldin Volcano on Unimak Island (Fournelle, 1988; Fournelle and Marsh, 1991). Additional mapping of Makushin Volcano on Unalaska Island (Nye and others, 1986) has also materially improved our geologic knowledge of this large and important volcanic center. Initial geologic and geochemical results from Fisher caldera have been reported (Fournelle, 1990b). Sampling for geochemical studies has also provided additional data for some previously mapped centers. This work includes that on northern Kanaga by Brophy (1990), on northern Adak by R. W. and S. M. Kay and coworkers (Kay and others, 1978, 1982; Kay and Kay, this volume) as well as Marsh (1976), and at Okmok Caldera on Umnak Island by Nye (1983) and Nye and Reid (1986).

HISTORY OF VOLCANISM

Episodes of volcanism and plutonism

Aleutian arc volcanism is not continuous, but episodic. Recent volcanism is recorded by ash layers in the adjacent sea-floor sediment (Hein and others, 1978; Scheidegger and others, 1980), as well as tephra layers in coal beds on the Kenai Peninsula (Reinink-Smith, 1990a). Hein and others (1978) showed that Aleutian volcanism has waxed every 2.5 m.y. for the past 10 m.y., and about every 5 m.y. for the previous 10 to 20 m.y. The present episode of strong volcanism began about 1 Ma, and an equally strong episode occurred about 7.5 Ma and perhaps another near 16 Ma. The chemistry of deep-sea ash samples (see section on Chemistry and Petrography) is skewed toward andesitic-dacitic compositions, so it is likely that basaltic volcanic activity is under represented.

The Kenai coal tephra layers preserve a more detailed (but less complete) eruptive record for the Alaska Peninsula segment of the arc. Reinink-Smith (1990a), using mineral radiometric dates and estimated peat accumulation rates, found a strong episode of volcanism more than 10.5 m.y. ago, with eruptions every 125 to 500 yr. A reduced level of activity (one eruption every 9,000 yr) occurred between 10.5 and 7.5 Ma, with a pulse at 7.5 Ma. Volcanic activity dramatically increased about 5 Ma, with events every 1,700 to 2,400 yr.

These magmatic episodes are also apparent in the subaerial geology, especially in the central Aleutian Islands, where erosion has been deep enough to expose plutons within the volcanic pile. On Adak, the initial products of the present phase of volcanism occur as erosional residuals of fairly fresh lava flows and domes respectively near Heart Lake and along Kuluk Bay (Coats, 1956). These units are ~3 to 3.5 m.y. old (Marsh, 1976), and, in contrast to younger volcanic flows, commonly contain small amounts of carbonate and zeolite. On Atka, quartz diorite (8 to 9 Ma) crops out at the headland between Crescent Bay and Bechevin Bay, whereas basaltic lava flows (~11 Ma) occur at Martin Harbor and Bechevin Bay (Hein and others, 1984). Plutons near Sergief Bay as well as on Kugalaska have ages of 12 to 14 Ma. Vallier and others (this volume) summarize the setting of these and other plutons and consider the rocks into which they intrude.

This long history of volcanism has produced three main sequences of sedimentary-volcanogenic rocks, which have been most extensively studied and defined on Adak, Atka, and Amlia Islands. The Finger Bay volcanics, defined by Coats (1947), represent the lower series (~55 Ma) on Adak. The middle series (~35 to 10 Ma) apparently is not common in the central arc, and the upper series (<10 Ma) represents the more recent volcanism.

Sedimentary and igneous rocks of the entire arc result from volcanic episodes that have yielded plutons, lavas, pyroclastics, and volcaniclastics. Former emplacement and depositional environments were similar to today's, having local, quiet, and deep basins in the oceanic sector and shallow, swampy tidal flats in the continental sector.

Historic volcanic activity

Recorded Aleutian volcanic activity dates from the beginning of the eighteenth century, according to sparse written accounts (Grewingk, 1850; Dall, 1870; Veniaminov, 1840; Becker, 1898). Dall (1918), various U.S. Geological Survey geologists (in Coats, 1950), and the *Volcano Letter* of the Hawaiian Volcano Observatory (e.g., Jagger, 1932) also provide fragmentary information. The historical record, however, is far from complete, owing to a small native population concentrated in a few villages on a handful of islands. Dumond (1979) and Black (1981) have described the effects of volcanism on the indigenous peoples of the Aleutian Islands and Alaska Peninsula.

Many volcanoes are obscured by cloud cover for much of the year, hindering observation. Today most information about eruptions comes from aircraft pilots and satellite observations (e.g., McClelland and others, 1989). Much of the uncertainty in the early reports of volcanic activity stems from observations that simply record "smoke," some of which could be summit clouds. Although reporting of volcanic activity in the arc has materially improved in recent decades, a significant amount of volcanic activity may still go unreported.

Coats (1950), Simkin and others (1981), and McClelland and others (1989) document the record of volcanic eruptions over the past 300 years. Reported eruptions since 1700 are shown in Figure 1. The most active volcanoes have been Pavlof (42 eruptions), Shishaldin (33), and Akutan (30). The most active volcanoes are those on the western end of the Alaska Peninsula and the easternmost islands of the Aleutian arc.

DISTRIBUTION OF VOLCANISM

Volcanic front: Segmentation and spacing

The most distinctive spatial feature of present Aleutian volcanism is that "In detail, the volcanic line does not form a perfectly simple arc, but consists of segments of different lengths . . ." (Coats, 1950, p. 45). Coats recognized that arc vol-

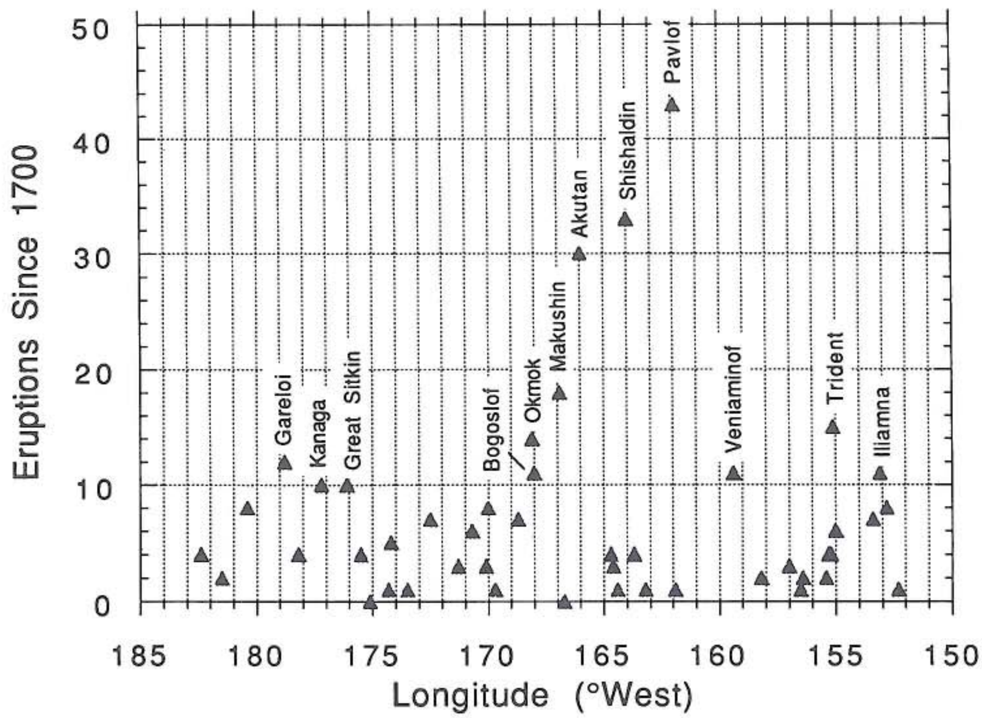


Figure 1. Plot of reported eruptions of Aleutian volcanoes since 1700 versus longitude.

canism produced a linear, segmented volcanic front. Marsh (1979a,b) used the active vents (i.e., summit craters) to more precisely define this segmentation, which is impressive when viewed from a volcanic summit, and holds best for the youngest volcanoes. Bogoslof Island, 35 km north of Umnak, and Amak Island, 50 km northwest of Cold Bay, are the only two places where subaerial volcanoes occur significantly off the volcanic front; they suggest a weak secondary volcanic front (Marsh, 1979a,b).

The intensity of the volcanic front, as measured by the magmatic flux (i.e., estimated subaerial volcano volume) over the most recent episode of volcanism, decreases strongly from east to west (Fig. 2). Also, the most active volcanoes over the past 300 years are located where convergence is the greatest (Fig. 1). The westernmost sign of recent subaerial volcanism is tiny Buldir Island near 176°E. This diminution of volcanism coincides with the systematic westward reduction in the rate of subduction of the Pacific Plate. The subduction rate in the far eastern part of the arc (152°W) is about 6.4 cm/yr (Minster and others, 1974), increases to about 7.1 cm/yr at 165°W, and then decreases continuously westward until 176°E, where the relative plate motion becomes strike-slip with no component of underthrusting.

The curvature of the arc is not the same as that of the earth itself, and so the subducting Pacific Plate is segmented into platelets or locally strongly bent. This deformation is indicated by variations in the record of seismicity (e.g., Spence, 1977; Kienle and Swanson, 1983). Major irregularities and breaks in the subducting plate mainly occur along original fracture zones (e.g., Amlia and Adak fracture zones) formed at a spreading center. The distinct change in the volcanic front east of Katmai (near Douglas at 153.5°W) correlates with a bend or kink in the subducting plate. There also may be some segmentation of the arc plate (e.g., Geist and others, 1988).

The position at depth of the subducting plate is fairly well known (e.g., Engdahl, 1973; Fujita and others, 1981; Plafker and others, Plate 12, this volume). As noted originally by Coats (1962) in his seminal introduction of the concept of subduction, the volcanic front is ~110 km above the seismic zone. The dip of the plate at these depths is generally about 45°, but at the shallowest part the dip decreases systematically eastward as the arc-trench gap widens; Jacob and others (1977) suggest that the Aleutian Trench in the Gulf of Alaska has migrated ~200 km seaward due to sediment accretion over the last 15 m.y. or so.

In the Aleutian oceanic sector, the long episodic history of Tertiary volcanism has produced extensive volcaniclastic deposits. These form the Aleutian Ridge, a large submarine basement structure. Most of the present landmass is thus related to much earlier phases of arc activity; the origin of the ridge itself is discussed in Vallier and others (this volume). An allochthonous terrane, the Umnak Plateau, forms the basement from west of Unimak Island to the Islands of Four Mountains (Howell and others, 1985).

The Alaska Peninsula has more complex basement of Early Permian and Mesozoic age, which is part of the Peninsular Terrane of Silberling and others (this volume). Wilson and others (1985) recognize two subterranes based on the age of the

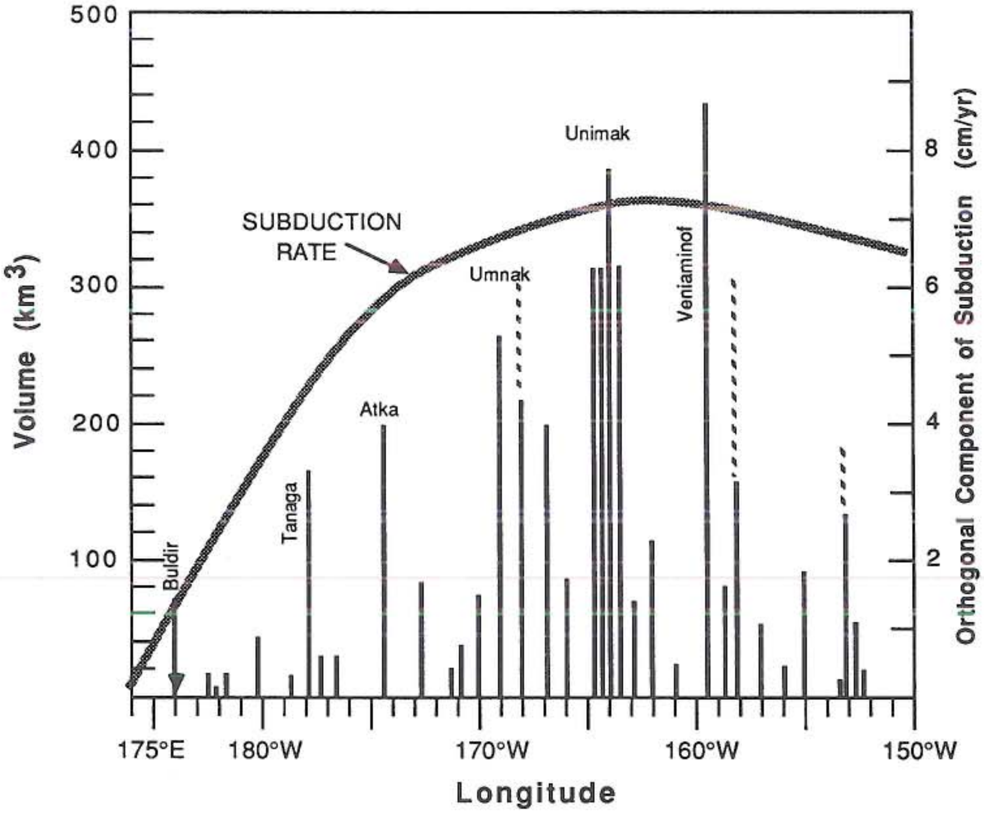


Figure 2. Estimated subaerial volumes of recent (≤2 Ma) Aleutian volcanic centers plotted in relation to their position along the arc. Volumes are calculated based upon the subaerial volcanic structures; they represent minimal volumes of extrusive material and are given to show the relative differences in sizes of volcanic centers. Dashed lines indicate uncertainties in calculating volumes. Also plotted is the orthogonal component of the rate of subduction of the North Pacific Plate along the arc (based on velocities of Minster and others, 1974). Note the correlation between volcano size (volume) and rate of subduction. (From Marsh, 1987.)

overlying Cenozoic volcanic rocks. The oldest subterrane occurs in the central Alaska Peninsula, formed during Eocene to early Miocene (48 to 22 Ma) volcanism (the Meshik arc of Wilson, 1985), is subparallel to the current one, and consists of basaltic to dacitic volcanic and intrusive rocks and intercalated volcaniclastic deposits. These rocks form the "transitional" crust upon which the presently active volcanic arc southwest of 157°W is built.

The present volcanic front in most instances coincides with earlier fronts. Most active centers in the volcanic front rest on the broad basement structure of the Aleutian Ridge, and Bogoslof and Amak Islands of the secondary front lie northward of this structure. Semisopochnoi Island appears to be an exception, where no older island-arc deposits exist.

Judging from the locations of plutons and the eroded remnants of earlier volcanoes, the volcanic front has moved systematically northward over the past 5 m.y. This migration is greatest in the centtral and western islands and lessens eastward. Midway up the Alaska Peninsula little, if any, migration is apparent. On Kiska, Tanaga, Kanaga, Adak, and Atka, volcanism currently occurs on the northernmost edges of the islands. The migration has not always been northward for, judging from the positions of the oldest plutons, volcanism may have first shifted slightly southward and then northward over the past 15 to 20 m.y. In some areas, like Tanaga, Seguam, and eastern Unimak, the migration may have been from east to west, parallel to the volcanic front. The magnitude of the northward motion on Adak and at Cold Bay may have been about 8 to 9 km over the past 3 to 5 m.y. Some of the largest volcanic centers (Okmok on Umnak and Veniaminof, Aniakchak, and others on the peninsula) and perhaps the Islands of Four Mountains may have not moved much during the past 1 m.y.

The continental sector (east of about 163°W) consists more commonly of large single stratocones, whereas in the oceanic sector the volcanic front is defined mainly by clusters of volcanoes, with each cluster defining a volcanic center. Regardless of size, active volcanic centers are usually spaced at intervals of about 60 to 70 km throughout the arc (Marsh and Carmichael, 1974), although the Katmai group of volcanoes appears to be an exception, having ~13 km intragroup spacing (Kienle and Swanson, 1983). Marsh (1979b) noted a rough correlation between spacing and volume, with the largest volcanic centers having a somewhat regular spacing. Some of these centers are presumably older, and younger centers have subdivided the original intervals. This process continues, as exemplified by the small and very young volcanoes of Bobrof between Tanaga and Kanaga, and by Kasatochi and Koniuji between Great Sitkin and Atka.

The segmentation of the front interrupts the regional spacing. Spacing within each segment may be independent of the next segment, making the intervals at the ends of segments irregular, for example, in the Islands of Four Mountains, near Douglas and Fourpeaked (~154°W), and the large interval between Atka and Seguam. There is no correlation between estimated subaerial volume and segmentation. Slightly more than half of the large centers lie within, rather than at the end, of their respective segments.

Tectonics and volcanism

S. M. Kay and others (1982) and Kay and Kay (this volume) proposed an alternative model of arc segmentation, in which local tectonic environment is the dominant control on magma chemistry. Their model (1) segmented the arc on a coarser basis, using islands rather than summit vents as the data points to be connected; (2) used the pattern of recent seismicity and tectonic (fracture zone) boundaries; and (3) gave special attention to 1957 and 1965 earthquake aftershock zones. The basis for assuming that these aftershock zones have been fixed throughout the current 1 to 2 m.y. of volcanism, however, is unclear. Based on these criteria, they divided the arc into four major tectonic blocks. They suggested that the size and composition of a volcanic center is determined by its position within a block. At the edges of blocks (extensional stress regimes), magmas ascend rapidly, experience little heat loss, and fractionate mostly at shallow levels producing large, tholeiitic centers. In the central parts of blocks (compressive stress regimes), magma ascent is inhibited, so that magmas ascend slowly, lose large amounts of heat rapidly, and fractionate at depth. By this model, small, calc-alkaline centers dominated by andesite are produced.

An alternative model for the generation of these two types of centers was suggested by Myers and others (1985). Based on Sr isotopic data from Atka and Adak, these authors suggested that the volumetric and geochemical differences (i.e., tholeiitic versus calc-alkaline) betwen the two volcanic centers could be explained by variations in the degree of development of the lithospheric magmatic conduits supplying each center.

The Aleutian Ridge in the central and western arc is fragmented into several structural blocks (Geist and others, 1987, 1988). At approximately 5 Ma, clockwise rotation of these blocks began, producing extended portions of the arc crust with summit basins. The modern volcanic centers of Kanaga, Adak, and Great Sitkin developed on unextended arc crust and Atka on incipiently extended crust, whereas Seguam grew atop strongly extended crust. Singer and Myers (1990) suggest that these contrasting crustal conditions may have had an effect on the different magmatic evolutionary histories of these volcanic centers.

EVOLUTION OF VOLCANIC CENTERS

Larger centers

Oceanic sector. Major volcanic centers along the arc show a remarkably similar sequence of morphologic development that varies significantly between the oceanic and continental parts (see Wood and Kienle, 1990, for brief descriptions with maps/photos of each Aleutian volcano). Myers and others (1985) classified Aleutian volcanic centers by their sizes (volumes), with large centers greater than 130 km^3 and small with less than 10 km^3.

In the oceanic sector, the lowest exposed subaerial lavas are mainly basaltic and form thin (3 to 5 m) flows in a gently dipping stack. These lavas are interspersed with coarse tephra and distributed around a central vent. The number of flows, which can vary from two or three to more than 20, are proportional to the size of the volcano. With continued eruption, the proportion of pyroclastic materials, especially lahars, increases greatly until only an occasional lava flow is found. Many units begin at the central vent, which is in most instances the locus of all eruptions. These flows soon encounter snow and ice or muddy debris, promoting autobrecciation, and rush down the volcano, sweeping the slopes as hot debris flows. These lahars, which may be as thick as 30 to 40 m, often show a multitude of detailed internal sedimentary structures, and fill glacial and river valleys, coming to rest with flat tops.

The deposition of coarse pyroclastic material, with a higher angle of repose than lava, results in the buildup of a steep stratocone upon the earlier shield. Abrupt changes in the style of the cone may reflect a change in the lava composition and/or viscosity of lavas that are erupted from a central vent amidst fields of ice, snow, and muddy debris.

During construction of the stratocone, the third major phase of volcanism begins with the formation of one or more new vents at some distance from the central vent on the flanks of the original shield. These smaller satellite cones, which may number as many as four or five, build simultaneously with the central cone and invariably show intercalated units. With continued buildup the compositions of the lavas of the main stratocone eventually become andesitic, then dacitic, whereas that of the satellite cones remain basaltic. Dome formation occurs in the dacitic stage of the main stratocone, but late plugs of the satellite vents remain basaltic. Near all the vents, small dikes and sometimes sills (10 to 100 cm wide) cut the volcanic pile.

Regardless of volcano size, the final cone-building stage is often culminated by caldera formation and eruption of a dacitic ash flow (Semisophochnoi), large dome (Atka), or rhyodacitic pumice beds (Okmok). The central-satellite vent systems cease eruptions and further activity is generally associated with the caldera itself. New activity consists of eruptions on the floor of the caldera, if the caldera floor is large and at low elevation (like Okmok, and in the continental sector, Aniakchak), or of growth of a new cone near the caldera rim, if the caldera is smaller and the floor is at high elevation (like Atka and Cold Bay). When eruptions within the caldera cease, activity shifts to a new location a few kilometers away where a new stratocone is built. Once established, eruptive activity in most instances stays within a relatively restricted circular zone about 10 to 20 km in diameter. Exceptions to this pattern occur on Tanaga, Seguam, and eastern Unimak, where the volcanism has migrated westward parallel to the arc some 10, 5 to 10, and 30 km, respectively, and in the continental sector at Pavlof where it has migrated about 20 km northeast.

Although high rates of erosion may occur during buildup, once dormancy or extinction comes, craters and high-level calderas become catchment basins for snow and ice builds. Basins are soon breached and become highly active cirques, excavating downward through the throat of the volcano and exposing the internal details of the vent.

Continental sector. In the continental part of the arc, major volcanic centers build with little sign of developing satellite vents, and there is more of a tendency to establish large solitary stratocones. Some have an array of substantial subsidiary cones, such as at Pavlof and adjoining Pavlof Sister, Little Pavlof, Double Crater, Mount Hague, and Mount Emmons, and perhaps at Katmai and Mageik. There is a tendency also for successive cones in any one region to form a linear series rather than a cluster as in the oceanic part of the arc. Also, the largest cones, beginning with Shishaldin, lie on or landward of the continental slope ($\sim 165°W$). The chemical composition of the lavas and pyroclastics also changes eastward along the arc. Basalts and basaltic andesites volumetrically dominate the oceanic sector of the arc, whereas on the Alaska Peninsula basalts are far less common, or even absent from centers such as Katmai. Instead, they give way to more siliceous compositions that occur as voluminous and abundant ash flows. This transition in composition, which probably occurs near Veniaminof (159°W), may also be reflected in postcaldera eruptions. Miller and Smith (1984) found that these eruptions east of Veniaminof continue to be silicic, whereas Veniaminof and centers westward return to more mafic compositions.

Smaller centers

There are many small (<10 km^3) volcanic centers, especially in the oceanic sector, that show another style of development. Some evolve much as a large center, but never establish satellite vents, and generally in the last stages of activity produce summit and flank domes. The summit domes are andesitic to dacitic and the flank domes more basaltic. Others are more akin to large domal centers and although they produce lavas, they never develop a well-defined central eruptive crater. Their extrusive materials are for the most part highly crystalline. In many ways these centers (such as Adagdak and Moffett on Adak) may represent a critical point: for any less magma, no volcano would occur, whereas for more magma, a full stratovolcano develops.

Cinder cones

Cinder or monogenetic cones are not common throughout the arc, but where they are found, they are plentiful. On Shishaldin, two dozen cinder cones are scattered over the northwesterly flank at low elevations ($<1,000$ m), and some have issued substantial lava flows (Fournelle, 1988). Westdahl and Pogromni are surrounded by at least 22 cinder cones (Neal and Swanson, 1983). Eleven recent cinder cones occur in the vicinity of Pavlof Volcano (Kennedy and Waldron, 1955). These are small with a maximum height of about 100 m, and are similar to ones that occur on the flanks of other volcanoes, especially near vents that have issued lavas. They have a subparallel alignment to the six major stratocones (McNutt and Jacob, 1986). In addition, a group of cinder cones stretches in a line northwesterly from the northern flank of Veniaminof (Burk, 1965). The sizes of these postglacial cones seem to diminish away from the volcano. Cinder cones are also found on the floors of large calderas at Okmok, Aniakchak, and Fisher. Several of these have issued substantial amounts of tephra and lava.

The paucity of cinder cones within the arc may reflect the dominantly central-vent style of eruption characteristic of stratocones. Only when the central vent becomes high and narrow, or otherwise congested, is it easier for magma to be dispersed by dikes. On the southwest flank of Atka Volcano, a regional dike that likely formed accompanying caldera formation has been glacially excavated; apparently it was the feeder dike for a cinder cone. Cinder cones suggest fissure-style eruptions. Historic fissure eruptions are known in the Aleutians: Gareloi was split from summit to flank in April of 1929 (Coats, 1959a) with the creation of 13 explosion craters; three fissures formed on the north slope of Shishaldin in 1830 (Veniaminov, 1840); and Seguam had a fissure eruption in 1977.

In proposing a method to map regional stress patterns, Nakamura (1977) and Nakamura and others (1980) used cinder cone distributions and dikes at several Aleutian centers as indicators of the direction of the maximum horizontal stress. The results mainly seem to reflect a smoothly varying stress field dictated by the direction of convergence. However, direct evidence of regional stress is scarce because there are few cinder cones and fissures. Areas with major indicators of stress that Nakamura and associates did not consider include the alignments of active vents on Tanaga and at Pavlof, fissures and vents on Seguam, a cinder cone field (and adjacent dikes) on Shishaldin, and a set of regional dikes on Atka. Tanaga and Pavlof indicate a northeast-southwest principal stress, whereas Atka, Seguam, and Shishaldin show a northwest-southeast stress. Together they suggest that the regional stress field is not uniform, but is influenced by local stresses (e.g., McNutt and Jacob, 1986).

GEOCHEMICAL AND PETROGRAPHIC FEATURES OF THE ALEUTIANS

Chemistry and petrography

The early works of Fenner (1926) in the Katmai region and Coats (1952) in the central Aleutian Islands suggested a calc-alkaline character of the modern volcanic rocks. As the number of volcanic centers sampled has increased, however, the Aleutian oceanic suite has been shown to have greater compositional diversity than originally believed.

A large and varied geochemical and isotopic database exists for the Aleutian arc. The number and types of analytical data available for individual volcanic centers, however, varies considerably. Of the 39 Aleutian island volcanic centers, geochemical and/or isotopic data are available for only 23 (Appendix 1). A significant number of these are characterized by only a few analyses of a few elements. The analytical data set is also skewed toward the large volcanic centers of Shishaldin, Makushin, Okmok, and Atka, plus smaller Seguam, which have large and comprehensive geochemical databases. For these centers, many samples have been fully characterized, having been analyzed for REE and isotopes as well as major and trace elements. Other volcanic centers have been analyzed for only a single geochemical variable (e.g., REE or isotopes), and so their geochemical characterization is fragmentary.

Many of the 30 islands with recent volcanic activity have multiple centers, and therefore the number of major individual volcanoes is much larger than the number of islands. The islands range in size from Unimak (~4,000 km^2) to Koniuji (<1 km^2). These geologic differences suggest that volcanoes are supplied by magmatic systems that vary considerably in degree of development and complexity. Large volcanic centers with many eruptive stages and satellite vents, for example, may have noncommunicating *crustal* reservoirs and long-lived magma chambers (e.g., Myers and Frost, 1989). Lavas from such plumbing systems probably record the influence of a variety of magmatic processes, for example, crystal fractionation/accumulation, magma mixing, and crustal assimilation. Smaller centers, on the other hand, may be supplied by smaller and simpler systems. The geologic complexity of Aleutian volcanic centers may also reflect differences in *subcrustal* magmatic processes. In view of probable lithospheric and crustal complexities, analysis of chemical compositions without corresponding geologic information is of limited use in interpreting the processes of arc-magma generation and evolution. Geochemical data should be compared and contrasted between volcanic centers whose geologic setting and evolution are similar. The geology, geochemistry, and petrology of individual Aleutian volcanic centers is reviewed by Myers (in preparation).

The following section summarizes the general chemical features of Aleutian volcanic rocks. Various representative compositions from the volcanic centers of Cold Bay, Shishaldin, and Atka are given in Table 2. Silica-oxide variation diagrams for these three centers are presented in Figures 3 to 5.

Rocks are divided by SiO_2 content: basalts, $SiO_2 < 53$ wt. %; andesites, 53 to <63 wt. %; dacites, 63 to 70 wt. %. There are only a handful of published analyses of rhyolite, reflecting mainly the general scarcity of this composition, but also perhaps the tendency to ignore ash layers in soil when sampling. Maximum SiO_2 content is 76 wt. % at Vsevidof (n = 1) and Kanaga (n = 1); values of 72 to 75 wt. % are reported for Unimak (n = 2) and Umnak (n = 3), and 70 to 71 wt. % for Seguam (n = 7). On the Alaska mainland, Novarupta has several analyses of 76 to 78 wt. % SiO_2, but otherwise, between Mount Spurr and Cold Bay, there are no other volcanic rocks known to contain between 68 and 76 wt. % SiO_2 (see Miller and Richter, this volume).

Average compositions for the entire arc have also been calculated (Table 3), to provide a rough measure of chemical composition of Aleutian basalts, andesites (further divided into basaltic and silicic andesites), and dacites. The sources of all data are given in Appendix 2.

Basalts. Myers (1988) evaluated published chemical data for Aleutian basalts (<52 wt. % SiO_2; n = 205), dividing them into three major classes; high-MgO (>9 wt. %), low-MgO (<6 wt. %), and transitional (6 to 9 wt. % MgO) basalts. The low-MgO basalts were subdivided into low- and high-alumina (HAB) groups; the first subgroup has <18 wt. % Al_2O_3, <450 ppm Sr, >1.0 wt. % TiO_2, and >9 wt. % FeO*. Myers (1988) found that high-MgO basalts constitute 10% of the analyzed Aleutian basalts and occur at only a few centers; transitional basalts constitute 22%, and low-MgO basalts constitute 67% of the basaltic analyses. All basalts have intermediate levels of K_2O (0.5 to 1.0 wt. %).

The data set reviewed here includes more basalts (n = 346), with the upper SiO_2 limit at 53 wt. %. These basalts are very similar to those discussed by Myers (1988), with a few differences. For example, Al_2O_3 shows a bimodal distribution, with maxima at 16 to 17, and 18.5 to 19.5 wt %; to some extent, this reflects the inclusion of the Shishaldin data with its distinct FeTi-enriched, low-Alumina (low-Al) basalt. Similarly, whereas most basalts have TiO_2 in the 0.6 to 1.2 wt. % range, there is a distinct subset with values between 2.0 and 2.5 wt. %. Also, Sr shows a single maximum of 400 to 550 ppm (albeit with a rather large deviation), rather than a bimodal distribution as suggested by Myers (1988). Ba and Rb are present at, respectively, 200 to 300 ppm and 10 ppm levels; most Zr contents are in the 40 to 90 ppm range, whereas the range of Y is restricted to 15 to 25 ppm. The major element with the widest range of abundances is MgO (Cr and Ni have similar large ranges).

A common characteristic of island-arc lavas is their low (relative to MORB) concentration of base metals such as Ni, Co, and Cr. Ni contents in the basalts are in most instances less than about 25 ppm, although 100 to 200 ppm Ni is present in high-MgO basalts. Similarly, Cr is normally less than 50 ppm, although it too can reach 300 to 600 ppm.

High-MgO basalts contain moderately abundant (<16-vol. %) Mg-rich (>Fo_{88}) olivine phenocrysts with small chromite inclusions, plus occasional clinopyroxene of diopside-salite composition (plagioclase phenocrysts are for the most part absent or else show disequilibrium features). These phases are set in a fine-grained groundmass of olivine, clinopyroxene, plagioclase, magnetite, and sparse glass. Almost invariably, rocks with elevated whole-rock Ni (>40 ppm) and Cr (>50 ppm) contents contain Mg-rich olivine with Cr-spinel inclusions.

Transitional basalts have generally higher crystallinities (17 to 57 vol. % phenocrysts); An_{70-90} plagioclase phenocrysts dominate (Myers, 1988). Also present are lesser amounts of Fo_{50-70}

TABLE 2: REPRESENTATIVE MAJOR- AND TRACE-ELEMENT CHEMISTRY OF BASALTS, ANDESITES, AND DACITES FROM THE ALEUTIAN VOLCANIC CENTERS OF COLD BAY, SHISHALDIN, AND ATKA*

Volc. Center	Cold Bay	Cold Bay	Cold Bay	Cold Bay	Cold Bay	Shishaldin	Shishaldin	Shishaldin	Shishaldin	Shishaldin
Description	Morzhovoi Hi-Al Bas.†	Morzhovoi Bas. And.†	Morzhovoi Andesite	Frosty Peak Andesite	Frosty Peak Andesite	Recent Lava Hi-Al Bas.†	Glac. Lava Hi-Mg Bas.†	Recent Lava FeTi Bas.†	Somma Andesite	Dacitic Pumice
Sample No.	CB42	CB30	CB39	CB74	CB51	SH5	SH101	SH9A	SH134	SH141
SiO_2	51.40	53.43	59.40	58.03	61.54	49.83	48.99	50.89	59.84	64.14
TiO_2	0.87	0.94	0.83	0.76	0.48	1.67	1.24	2.21	1.07	0.72
Al_2O_3	20.40	19.25	17.53	18.37	17.66	20.29	15.11	15.73	16.74	15.97
Fe_2O_3	3.03	3.45	2.87	3.24	n.d.	3.32	2.69	3.04	2.21	0.70
FeO	5.52	5.14	4.38	4.07	n.d.	6.71	7.94	8.64	5.07	4.24
FeO*	8.25	8.25	6.96	6.99	5.90	9.70	10.36	11.38	7.06	4.87
MnO	0.17	0.18	0.16	0.16	0.12	0.18	0.18	0.23	0.18	0.22
MgO	4.04	3.26	2.94	2.71	2.26	3.62	10.56	3.99	1.93	0.99
CaO	9.95	8.41	6.76	7.51	6.46	10.49	10.75	7.83	5.30	2.73
Na_2O	2.67	3.34	3.53	3.60	3.73	3.20	2.16	3.77	4.51	5.54
K_2O	0.60	1.06	1.58	1.26	1.08	0.56	0.74	1.14	1.63	2.17
P_2O_5	0.28	0.30	0.30	0.24	n.d.	0.27	0.19	0.59	0.37	0.19
LOI	0.43	0.55	0.01	0.11	n.d.	0.72	0.80	0.91	1.40	n.d.
Ba	220	472	485	475	480	231	246	456	550	878
Co	28	n.d.	n.d.	n.d.	n.d.	27	53	27	23	n.d.
Cr	6	n.d.	n.d.	n.d.	3	22	434	17	5	23
Ni	10	n.d.	n.d.	n.d.	3	12	136	10	7	11
Rb	10	24	41	32	32	17	17	41	42	56
Sc	n.d.	n.d.	n.d.	n.d.	n.d.	27	41.1	30.7	18	n.d.
Sr	384	507	328	400	400	629	530	470	429	255
V	192	n.d.	n.d.	n.d.	76	240	316	181	64	28
Y	16	n.d.	n.d.	n.d.	19	23	17	43	40	46
Zr	68	125	179	142	127	76	85	215	246	263

Volc. Center	Atka	Atka	Atka	Atka	Atka	Atka	Atka
Description	Four Flows Hi-Al Bas.†	Potainkof Bas. And.†	Milky Valley Andesite	Konia Andesite	Korovin Andesite	Kliuchef Dacite	Glassy Dacite
Sample No.	AT 1	AT 33	AT 34	AT119	AT 117	AT118	AT 20
SiO_2	49.80	53.29	58.24	61.39	62.49	66.03	66.88
TiO_2	0.96	0.95	0.69	0.87	0.62	0.46	0.46
Al_2O_3	20.12	18.31	17.33	16.75	16.37	16.12	15.57
Fe_2O_3	4.36	3.83	3.58	2.71	1.99	1.69	1.25
FeO	5.51	4.71	3.02	4.13	3.95	2.64	2.52
FeO*	9.43	8.16	6.24	6.57	5.74	4.16	3.64
MnO	0.18	0.20	0.16	0.18	0.15	0.13	0.10
MgO	4.72	3.92	3.24	1.64	1.69	0.64	0.74
CaO	9.66	8.82	7.10	4.37	4.23	2.36	2.22
Na_2O	3.71	3.29	3.74	4.59	4.53	4.72	4.96
K_2O	0.78	1.11	3.74	2.79	3.00	4.60	4.09
P_2O_5	0.23	0.26	0.19	0.40	0.19	0.15	0.12
LOI	0.24	1.26	0.44	0.59	0.21	1.05	0.94
Ba	273	530	745	750	n.d.	890	862
Co	29	29	24	n.d.	n.d.	n.d.	38
Cr	42	29	50	3	n.d.	<1	18
Ni	20	7	11	<1	n.d.	<1	3
Rb	10	21	40	64	n.d.	92	116
Sc	n.d.	n.d.	18	20	n.d.	13	11
Sr	630	575	487	343	n.d.	178	155
V	275	n.d.	136	39	n.d.	21	20
Y	22	23	22	40	n.d.	44	40
Zr	64	99	128	257	n.d.	344	342

*Data sources: Cold Bay: Brophy, 1984, 1986a, 1987; Shishaldin: Fournelle, 1988; Fournelle and Marsh, 1991; Atka: Marsh, 1982a, unpublished data.

Oxide and LOI values are in weight percent (wt %); trace elements are in parts per million (ppm); FeO* = all iron as FeO.

†Hi-Al, high alumina; Bas., basalt; And., andesite; Hi-Mg, high magnesium; n.d., no data.

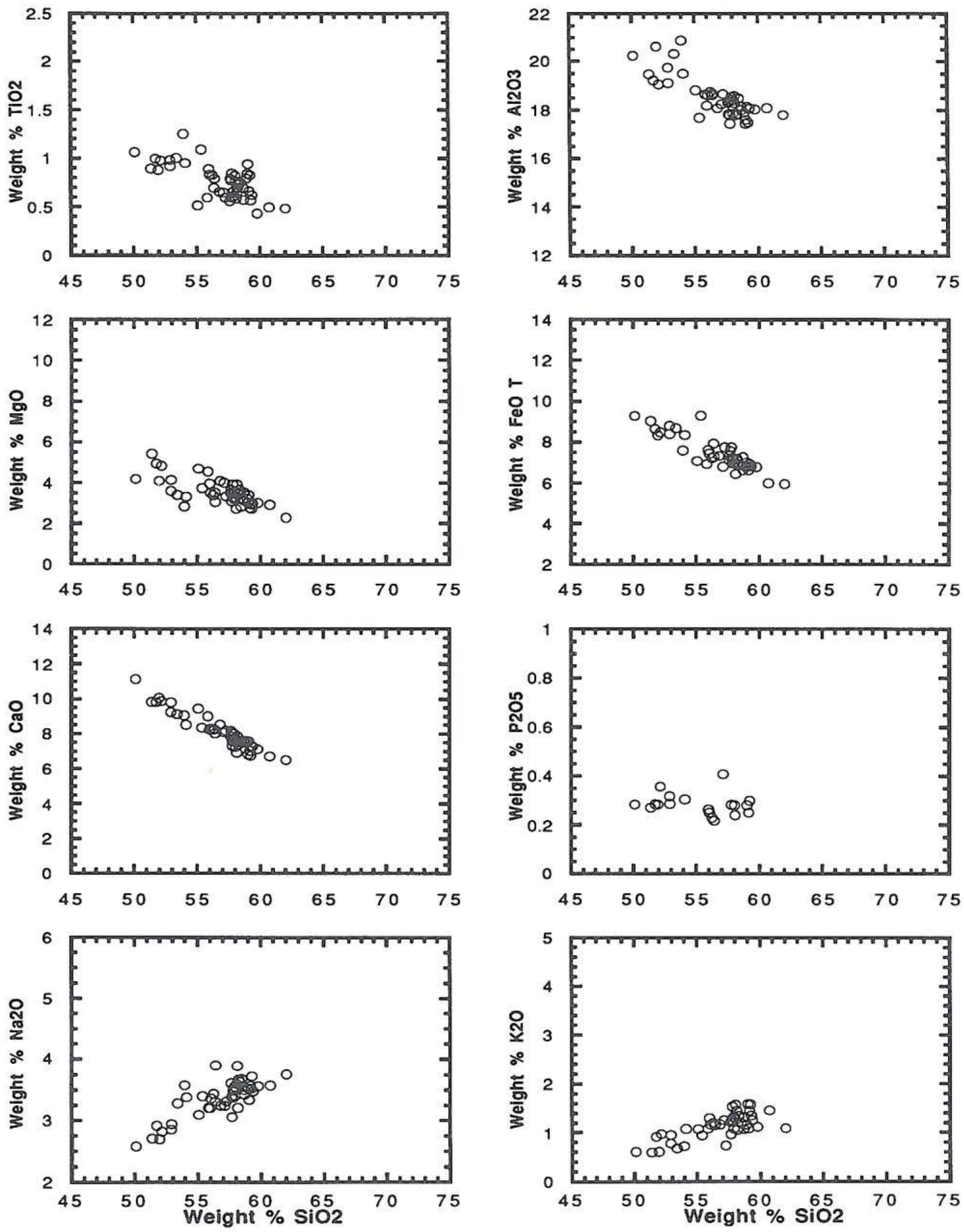


Figure 3. SiO_2-oxide variation diagrams for the Cold Bay volcanic complex. Oxides have been normalized to 100 wt. % "water-free" (samples with H_2O or LOI >2.5 wt. % not used). FeO T = all iron as FeO. Data sources given in Appendix 2.

olivine (3 to 7 vol. %), clinopyroxene (2 to 16 vol. %), and titanomagnetite.

High-alumina basalts (HABs) are highly porphyritic (25 to 73 vol. %), having abundant An_{70-90} plagioclase, lesser amounts of iron-rich (<Fo_{70}) olivine and titanomagnetite, and sporadic clinopyroxene (Myers, 1988). Groundmass phases in the low-MgO basalts include plagioclase, olivine, clinopyroxene, titanomagnetite, and rare interstitial glass. Pigeonite and hydrous phase phenocrysts are absent.

Little has been published on Aleutian low-alumina, low-Mg basalts. Myers (1988) recognized them as volumetrically minor. The FeTi-enriched, low-Al basalts from Shishaldin have, for the most part, less than 20% phenocrysts and in many cases, only microphenocrysts (Fournelle, 1988). Plagioclase dominates, with

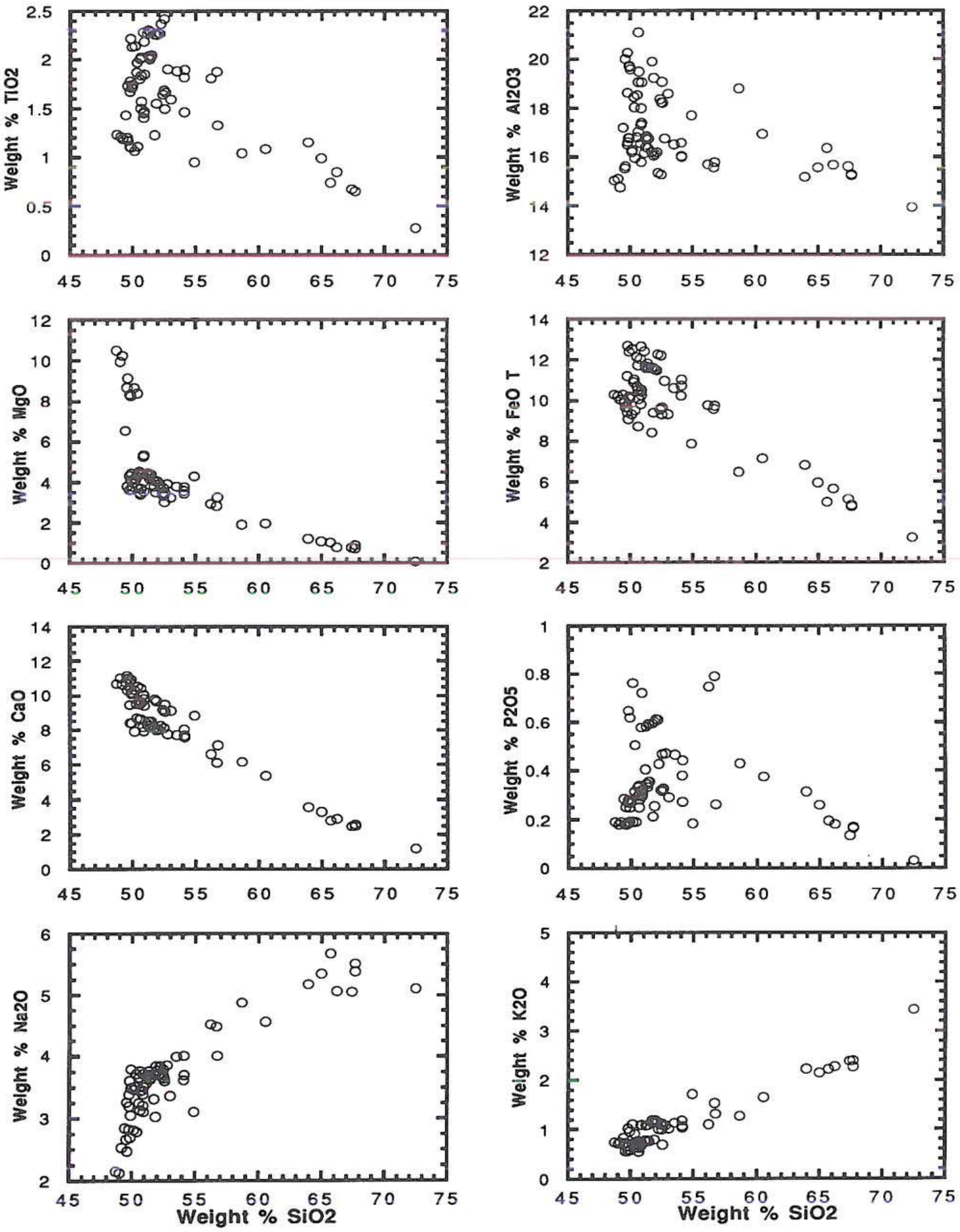


Figure 4. SiO_2-oxide variation diagrams for Shishaldin Volcano. Oxides have been normalized to 100 wt. % "water-free" (samples with H_2O or LOI > 2.5 wt. % not used). FeO T = all iron as FeO. Data sources given in Appendix 2.

$An_{50\text{-}70}$ microphenocrysts, and less common phenocrysts as calcic as An_{90}. Olivine (Fo_{70}) and titanomagnetite (ti-mt) are scarce (<1 vol. %), and clinopyroxene (cpx) is commonly absent; plagioclase, cpx, and ti-mt compose the groundmass. A few quench crystals of plagioclase and olivine are present.

Andesites. The Aleutian data set contains 594 samples with SiO_2 contents between 53 and 63 wt. %. Relative to the basalts, the andesites contain decreased levels of FeO* (6 to 8.5 wt. %), MgO (2.5 to 4 wt. %), and CaO (6 to 9 wt. %); higher Na_2O (3 to 4 wt. %) and K_2O (0.5 to 2 wt. %); and similar TiO_2 (0.6 to 1.2 wt. %) and Al_2O_3 (16.5 to 19 wt. %).

Most trace elements vary, with increasing silica, in a consistent manner. Ni and Cr are less abundant (<15 ppm and <30 ppm, respectively), although there are a few high values (~100 and ~300 ppm). Ba increases (300 to 500 ppm) as does Rb (15 to 45 ppm). Sr, however, remains relatively constant (300 to 600 ppm) within the basalt to andesite range.

The andesites have been further broken down into basaltic

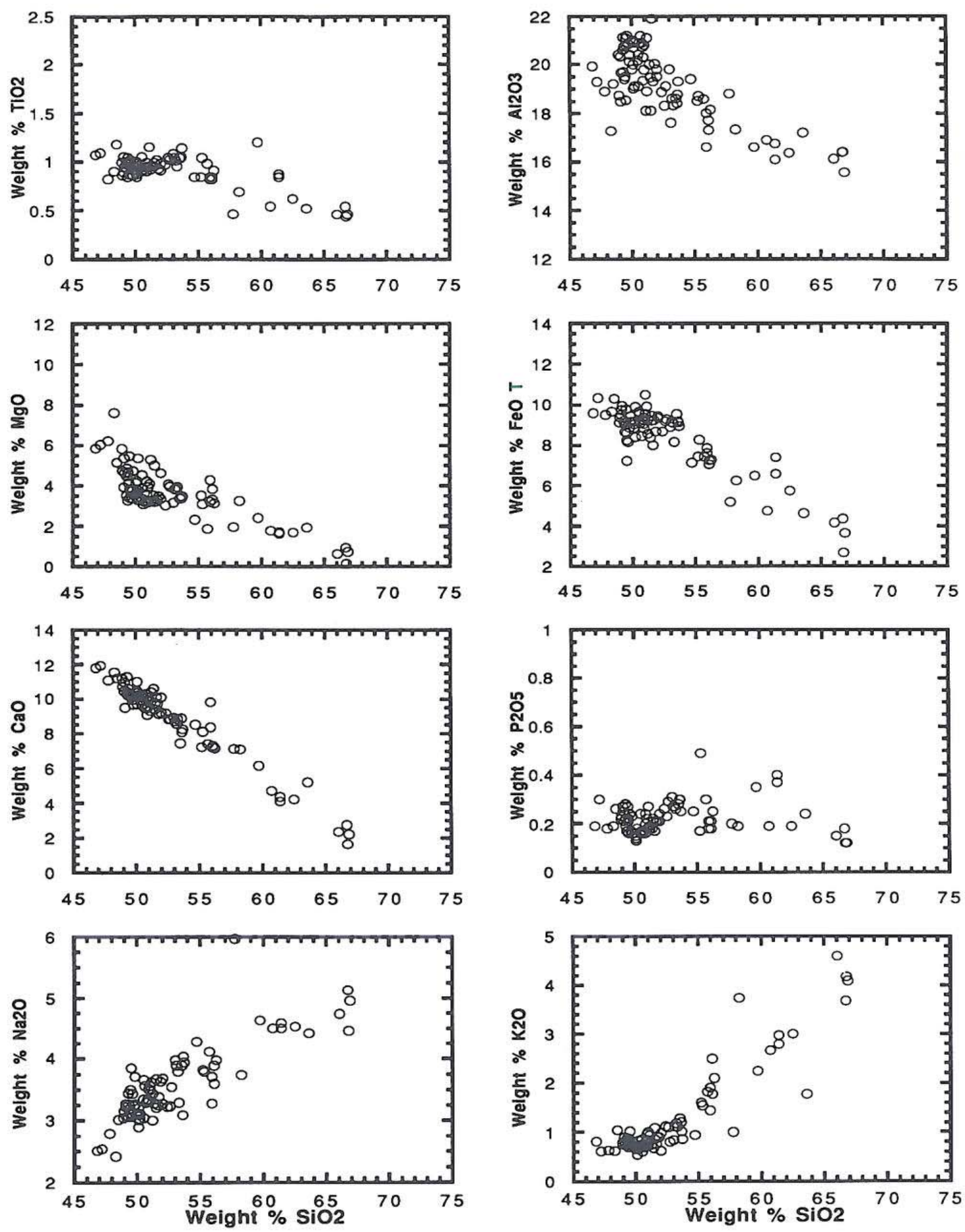


Figure 5. SiO_2-oxide variation diagrams for the Atka volcanic center. Oxides have been normalized to 100 wt. % "water-free" (sample with H_2O or LOI > 2.5 wt. % not used). FeO T = all iron as FeO. Data sources given in Appendix 2.

andesites (53 to 57 wt. % SiO_2) and silicic (acid) andesites (57 to 63 wt. % SiO_2). Calculated mean compositions are given in Table 3. The mean andesite composition is similar to the average worldwide andesite composition (n = 2,500) found by Gill (1981).

Andesites have the same high crystallinity as the basalts, approximately 35 to 50 vol. %. Plagioclase dominates, with 20 to 40 vol. %, followed by 6 to 9 vol. % cpx. Up to 3 vol. % orthopyroxene may be present; olivine and amphibole are absent or present in only trace amounts. Opaque minerals (mainly titanomagnetite) are more abundant (1 to 2 vol. %), at the volcanic centers such as Cold Bay, Akutan, Atka, Adak, and Kanaga, whereas at Okmok, Makushin, and Shishaldin they are generally absent.

Studies of andesites from the volcanic centers of Cold Bay (Brophy, 1987) and Kanaga (Brophy, 1990) show the following features: Kanaga plagioclase crystals range from fresh to corroded and sieve-textured, and the latter are more albitic than the former; most phenocryst cores are $An_{50\text{-}60}$. Cold Bay plagioclase show a bimodal distribution (An_{50-60}, An_{80-90}) and often complex zonation. Unzoned olivine (Fo_{62} to Fo_{81}, with ragged edges) is present in only one Cold Bay lava series; at Kanaga, Fo_{70-88} grains are corroded. Kanaga orthopyroxene compositions are in the range $Wo_{2-4}En_{58-71}Fs_{26-38}$, with the Mg-rich varieties showing normal zoning and the Fe-rich crystals reverse zoning (clinopyroxenes show the same patterns). Cold Bay clinopyroxene is in the range $Wo_{38-46}En_{40-47}Fs_{13-19}$. Amphibole is absent at Cold Bay, whereas trace amounts are present in one-third of all published Aleutian modes.

Dacites. A smaller number (n = 130) of dacites (i.e., 63 to 70 wt. % SiO_2) have been chemically analyzed. Compared to the andesites, there are significant decreases in FeO* (4.5 to 6 wt. %), MgO (0.5 to 2.5 wt. %), and CaO (2 to 6 wt. %), slight declines in TiO_2 (0.4 to 1 wt. %) and Al_2O_3 (15.5 to 16 wt. %), and increases in Na_2O (3.5 to 5 wt. %) and K_2O (1 to 2.5 wt. %). The major element with the largest spread of values is CaO (Sr has a similar wide range). The average Aleutian dacite composition is given in Table 3.

Ni and Cr (both <10 ppm) continue to decrease in concentration with increasing silica content. Ba and Rb increase respectively to 800 to 900 ppm and 30 to 60 ppm, whereas Sr decreases to 150 to 350 ppm. Zr shows elevated levels (125 to 275 ppm), and Y increases to 20 to 60 ppm.

Modal abundances have been determined for only a handful of samples (n = 8). These dacites have lower crystallinities, normally less than 20 vol. %. Plagioclase (≤15 vol. %) is the main phase, followed by cpx (<2 vol. %) and the opaques (≤1 vol. %). Small amounts (<1 vol. %) of olivine, orthopyroxene, and amphibole have been reported in individual samples but otherwise these phases appear to be rare. Plagioclase in Shishaldin dacites (Fournelle, 1988) ranges from An_{30} to An_{40}, green clinopyroxene is $Wo_{38}En_{36}Fs_{26}$, and golden-brown orthopyroxene is $Wo_4En_{49}Fs_{48}$. Kay and Kay (1985) report Great Sitkin clinopyroxene as $Wo_{42}En_{42}Fs_{16}$ and Westdahl clinopyroxene as $Wo_{40-41}En_{36-42}Fs_{18-22}$.

Hildreth (1983) determined mineral abundances of the 1912 Katmai (Novarupta) andesites, dacites, and rhyolites by mineral separation. He found the rhyolites to have 0.5 to 2 wt. % phenocrysts, whereas andesites and dacites contain 30 to 45 % (with dacites more crystal-rich). Present, in order of appearance, were plagioclase, hypersthene, Ti-magnetite, ilmenite, apatite, and pyrrhotite. In the rhyolites, quartz was as abundant as plagioclase; in the intermediate rocks, augite was about half as abundant as hypersthene.

The Deep Sea Drilling Project's (DSDP) northern Pacific cores, from locations adjacent to the Aleutian arc, contain volcanic ashes (Scheidegger and others, 1980). The ashes, of bulk

TABLE 3: MEAN COMPOSITIONS OF PUBLISHED ALEUTIAN VOLCANIC SAMPLES, FOR CATEGORIES BASED ON SiO_2 WEIGHT PERCENT, "WATER FREE"*

	Basalt		Basaltic Andesite		High-Silica Andesite		Andesite		Dacite	
Range	<53 wt% SiO_2		53-57 wt% SiO_2		57-63 wt% SiO_2		53-63 wt% SiO_2		63-70 wt% SiO_2	
	Mean (S.D.)	n=	Mean (S.D.)	n=	Mean (S.D.)	n=	Mean(S.D.)	n=	Mean (S.D.)	n=
SiO_2	50.6 (1.5)	346	54.9 (1.1)	270	59.4 (1.7)	324	57.4 (2.7)	594	65.8 (2.1)	130
TiO_2	1.10 (.42)	346	0.98 (.28)	270	0.80 (.21)	324	0.88 (.26)	594	0.72 (.17)	130
Al_2O_3	18.3 (1.9)	346	18.0 (1.1)	270	17.3 (.88)	324	17.6 (1.1)	594	15.7 (.71)	130
FeO*	9.57 (1.22)	346	8.17 (1.02)	270	6.68 (.96)	324	7.36 (1.24)	594	4.95 (.90)	130
MnO	0.18 (.03)	346	0.17 (.03)	269	0.15 (.03)	323	0.16 (.01)	593	0.14 (.04)	128
MgO	5.59 (2.25)	346	4.19 (1.02)	270	3.09 (.90)	324	3.59 (1.10)	594	1.53 (.65)	130
CaO	10.3 (1.2)	346	8.56 (.89)	270	6.79 (.89)	324	7.59 (1.29)	594	4.07 (1.10)	130
Na_2O	2.98 (.53)	346	3.43 (.45)	270	3.80 (.57)	324	3.63 (.55)	594	4.72 (.77)	130
K_2O	0.80 (.28)	346	1.15 (.36)	270	1.53 (.46)	324	1.36 (.46)	594	2.00 (.61)	130
P_2O_5	0.23 (.14)	322	0.23 (.11)	253	0.21 (.08)	274	0.22 (.01)	594	0.18 (.06)	128
Ba	309 (114)	193	401 (148)	110	584 (161)	102	489 (179)	212	710 (135)	74
Co	34 (9)	62	30 (8)	46	25 (8)	37	28 (8)	83	23 (12)	53
Cr	71 (133)	212	43 (32)	97	33 (72)	81	38 (54)	182	17 (13)	67
Cu	121 (75)	56	80 (118)	41	45 (36)	30	65 (94)	71	31 (43)	9
Ni	34 (61)	206	21 (25)	110	15 (19)	82	19 (23)	192	7 (6)	23
Rb	15 (8)	248	25 (10)	174	38 (12)	163	31 (13)	337	51 (18)	87
Sc	30 (7)	53	28 (8)	15	17 (6)	26	21 (7)	40	16 (3)	8
Sr	511	262	473 (140)	179	431 (135)	164	453 (139)	343	263 (79)	88
V	278 (66)	158	218 (55)	56	137 (70)	57	177 (75)	113	51 (44)	18
Y	24 (10)	213	24 (9)	159	29 (11)	137	26 (11)	296	39 (11)	82
Zn	82 (8)	17	74 (17)	6	84 (23)	4	78 (19)	10	81 (4)	2
Zr	88 (64)	222	102 (40)	172	136 (43)	145	118 (45)	317	180 (51)	82

S.D., standard deviation; n, number of chemical analyses. Oxide and LOI values are in weight percent (wt.%); trace element values are in parts per million (ppm); FeO = all iron as FeO.

SiO_2 54 to 65 wt. %, have small amounts of biotite and amphibole, with amphibole abundance increasing significantly for ash SiO_2 contents >66 wt. %. Biotite is uncommon in Aleutian subaerial deposits, and the only occurrence reported is in a rhyolitic dome at Recheschnoi (Byers, 1959). This rhyolite (75.4 wt. % SiO_2) contains "sparse, small euhedral phenocrysts of bluish gray quartz, feldspar, biotite, hornblende, and rare hypersthene" (Byers, 1959, p. 301).

Reinink-Smith (1990b) found that unaltered volcanic glass in the Pliocene coal beds of the Kenai Peninsula ranged in composition from andesite (60.3 wt. % SiO_2) to rhyolite (72.3 wt. % SiO_2).

Marsh (1982a) suggested two distinct styles of K_2O versus SiO_2 variation, leading to either ~2 or ~4 wt. % K_2O in dacites. The high potash trend is principally defined by the lavas of Semisopochnoi and Atka; both show a similar development from a central-satellite cone system to caldera formation and eruption of K_2O-rich dacite. High potash trends have also been found in some Japanese lavas (Aramaki and Ui, 1983).

Aleutian basalts and andesites have whole-rock sulfur contents in the range from 200 to 400 ppm (Marsh, unpublished data). Some volcanoes have large sulfur deposits, such as Little Sitkin, with up to 200,000 tons (Snyder, 1959), and Makushin, with up to 77,000 tons (Maddren, 1919). Ash erupted from Mount Spurr in 1953 was found to contain 1.5 wt. % SO_3 (Wilcox, 1959). Gerlach and others (1990) examined products of the 12/15/89 eruption of Redoubt Volcano and found up to 870 ppm of sulfur in melt inclusions in hornblende, 400 to 60 ppm in plagioclase and orthopyroxene melt inclusions, and ~60 ppm in the pumice matrix glass.

Arcwide variations in chemistry

Lavas and pyroclastics from Aleutian volcanoes compositionally span the SiO_2 range from 45 to 77 wt. %. Marsh (1982a) suggested that the dominant rock type in the arc was basaltic (49 to 52 wt. % SiO_2), based upon examination of the volcanic center of Atka. Examination of the numbers of published analyses for Atka and seven other Aleutian volcanic centers (Fig. 6) suggests that the situation may be more complex. The three very large volcanic centers of Shishaldin (300 to 400 km^3), Okmok (200 to 300 km^3), and Atka (200 km^3) are dominated by basalt, with minor amounts of dacite and andesite. The somewhat smaller volcanic centers of Makushin (145 km^3) and Akutan (80 km^3) appear to have approximately equal amounts of andesite and basalt, whereas Seguam (75 km^3) has a distinctive bimodal population of basalt/basaltic-andesite and dacite/rhyodacite. The small centers of Kanaga (25 km^3) and Adak (25 km^3, not shown on Fig. 6) are dominantly andesite, as is the volcanic compex of Cold Bay (~150 km^3) on the Alaska Peninsula. In addition to variations due to geographic position (continental versus oceanic, plus along-arc position), the volume of the volcanic center may be an important factor in determining its general chemical composition.

Figure 7 shows the longitudal variation of SiO_2 content of analyzed samples, from Mount Spurr in the east to Buldir Volcano in the west. Few basalts are present in the continental sector (i.e., east of Cold Bay); andesites through dacites predominate. Date from the oceanic volcanoes are presented here principally for comparison with those of the Alaska Peninsula–Cook Inlet sector.

Four oxides (Al_2O_3, FeO*, MgO, and K_2O) have been plotted against longitude, by silicia categories (basalts, basaltic andesites, silicic andesites, and dacites), in Figure 8a to d. Considering only the well-sampled centers (at a given SiO_2 level), the variability of chemical compositions at some centers is striking, compared to the rather restricted compositions at others. For basaltic compositions, Shishaldin and Okmok (and to some extent, Akutan and Makushin) show significant heterogeneity, particularly for TiO_2, Al_2O_3, FeO*, MgO, Y, and Zr. Atka and Seguam, on the other hand, appear to be homogeneous.

The andesitic database is relatively large and includes analyses of a number of centers, from Spurr to Yantarni, in the Alaska Peninsula–Cook Island section. Andesites from these volcanoes show a more restricted range (albeit for a small number of samples), relative to the oceanic sector, for Al_2O_3, TiO_2, FeO*, CaO, P_2O_5, Y, Zr, and, to some extent, K_2O.

Dacitic compositions are also shown in Figure 8. Dacites (and less common rhyolite, n = 12) are not restricted to the continental portion of the Aleutian arc. There appears to be a wider range of some oxides (e.g., FeO*, MgO, K_2O) in the oceanic sector compared with the Alaska Peninsula (AP)–Cook Inlet (CI) sector. In addition, K_2O appears to be somewhat enriched in the oceanic dacites and rhyolites relative to those in the AP-CI sector. For four oceanic rhyolites (72.5 to 76.4 wt. % SiO_2), there is an average of 3.9 wt. % K_2O, versus an average of 3.2 wt. % for four Novarupta rhyolites (which are at a higher silica content, ~77.3 wt. % SiO_2).

This arcwide variation in major (and trace element) chemistry is striking in the degree of diversity at some centers relative to others. There also appears to be less variation in composition of Alaska Peninsula–Cook Inlet basaltic andesites and silicic andesites, particularly in Al_2O_3, FeO*, and K_2O, compared to those present in the oceanic part of the arc.

Transverse to the arc the only measure of change is recorded by the volcanic rocks of Bogoslof and Amak. Bogoslof's heterogeneous rocks (basalts and silicic andesites) are more alkalic than those of the nearby volcanic front (Arculus and others, 1977; Marsh and Leitz, 1979). They are also lower in MgO, and their Sr isotopic values are, for the most part, lower.

Lavas from Amak consist of homogeneous andesite that is compositionally distinct (i.e., having lower MgO, and higher K_2O and Na_2O) from the lavas of the nearby Cold Bay volcanic center. Additionally, Amak Sr and Nd isotope values are distinct from those in Cold Bay lavas (Morris and Hart, 1983; Marsh and Lietz, 1979; von Drach and others, 1986).

The largest Aleutian isotopic database is for $^{87}Sr/^{86}Sr$ (n = 149). Values for the oceanic sector range from about 0.7028

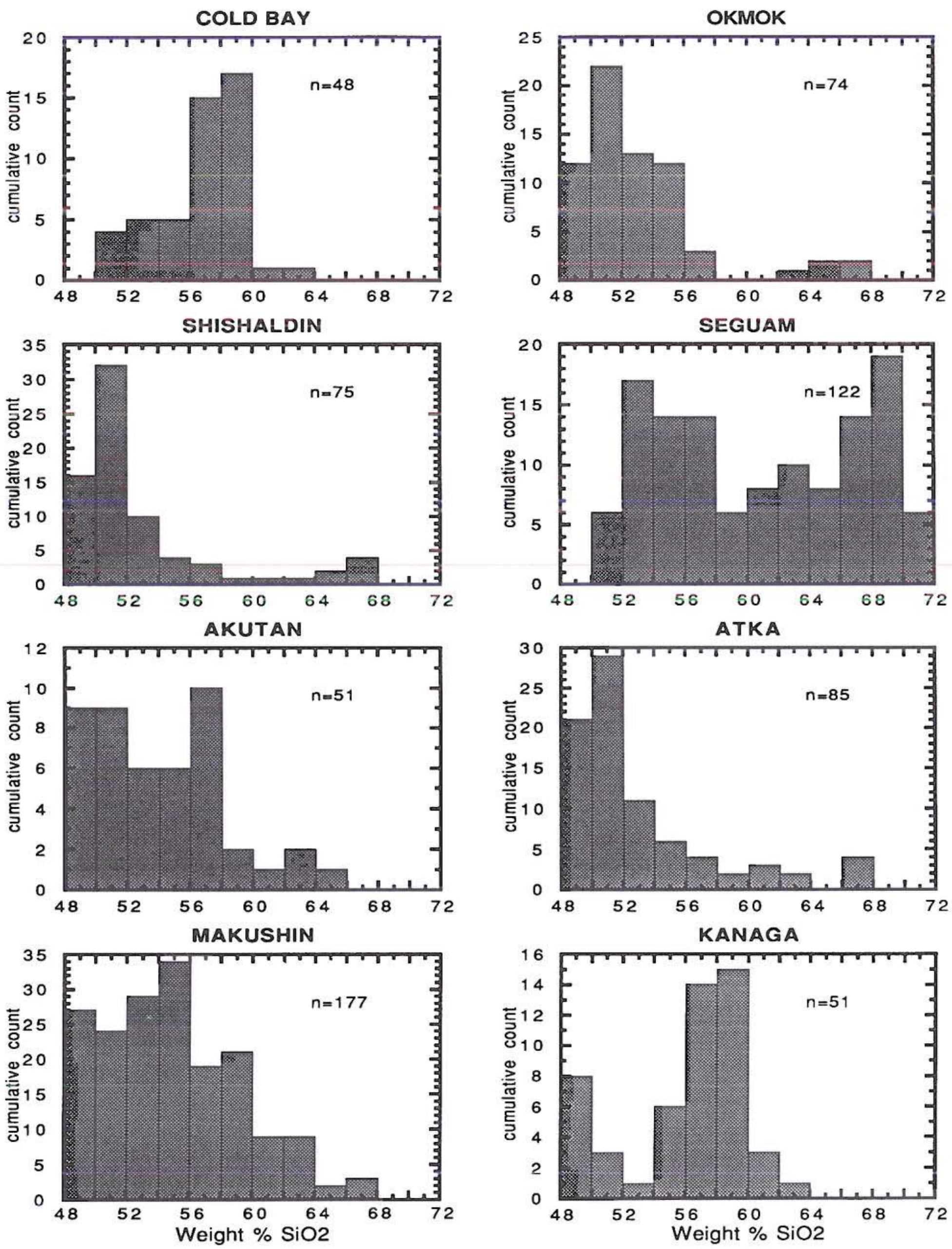


Figure 6. Histograms of number of published analyses versus SiO_2 (wt. %) for the volcanic centers of Cold Bay, Shishaldin, Akutan, Makushin, Okmok, Seguam, Atka, and Kanaga. Data sources given in Appendix 2. n = number of chemical analyses.

at Adak (sample ADK53 of Kay 1978, which we interpret as xenocryst-bearing) to 0.7037 at Seguam, with most data clustering between 0.70315 and 0.70335 (mean = 0.70331 ± 0.00024; Fig. 9). The coverage, however, is extremely variable, with one-third of the data from Adak and Atka. The secondary front volcanoes of Amak and Bogoslof have lower values, 0.7028 to 0.7032. With few data from the continental sector, there is a decrease in values from Spurr to Veniaminof, and continuing westward, average values are relatively constant although the least radiogenic values decrease to ~0.7029 through 167°W (Makushin). $^{87}Sr/^{86}Sr$ values show a distinct radiogenic enrichment at Seguam, with a westward decrease at Atka and then Adak.

Within this limited Sr isotopic range, a distinct feature is the

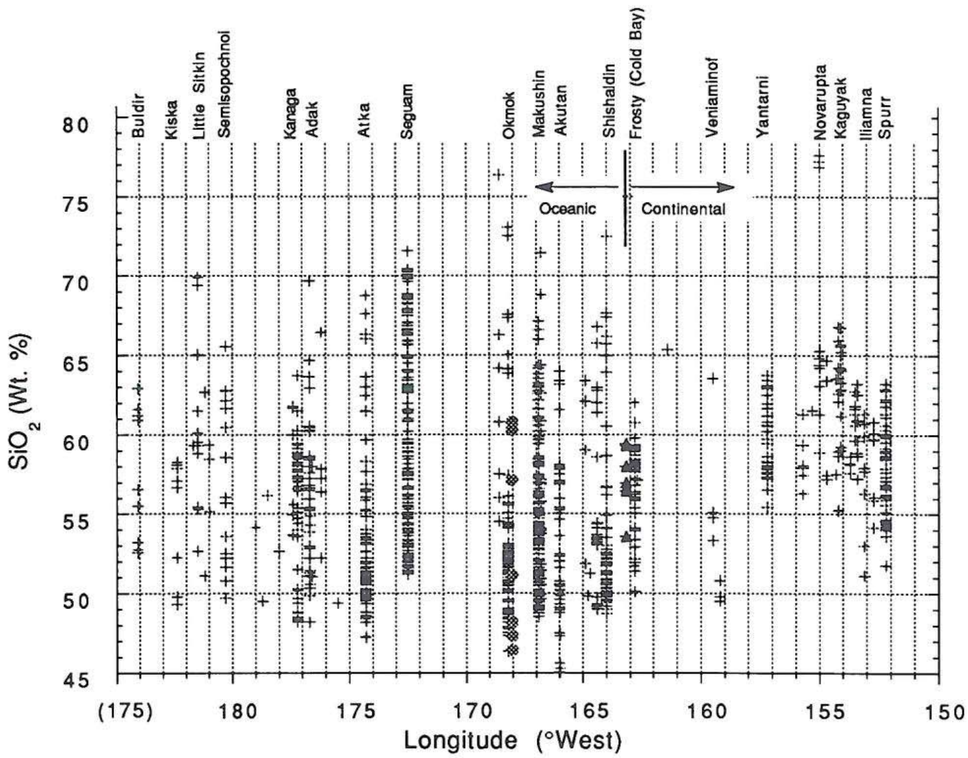


Figure 7. Plot of SiO_2 (wt. %, "water-free") versus longitude, from 152°W (Spurr) to 176°E (Buldir). Also plotted are the second-arc volcanoes of Amak (triangles; 163.15°W) and Bogoslof (hachured circles; 168.03°W). Data sources given in Appendix 2.

relatively large variation in some centers (for $n \geqslant 12$) such as Cold Bay, Okmok, and Adak, and the small variations at others, such as Seguam and Atka. Whereas this difference seems to reflect the diversity of lavas at some centers, such as Adak or Okmok, for some centers like Seguam, $^{87}Sr/^{86}Sr$ values show very little variation, despite the wide range of lava types present (basalt through rhyodacite).

The amount of $^{143}Nd/^{144}Nd$ data is limited (n = 68). Values range from 0.51294 to 0.51310 (Fig. 9), with an anomalous value of 0.51324 for Adak sample ADK53 (Kay, 1978). There is a range of data for each volcanic center, but there are only 7 to 9 data per center (12 for Seguam), making interpretation difficult. Seguam has the lowest range of values (0.51294 to 0.51299), similar to those for Cold Bay. There is a trend, from Seguam through Atka and Adak, of increasing $^{143}Nd/^{144}Nd$.

Lead isotopic data are limited (n = 80), with 75 percent from four centers (Okmok, Seguam, Atka, and Adak). There are differences in the spread of Pb isotopic values between these four centers (Fig. 9). Adak (n = 21) and Okmok (n = 15) show the greatest range of values, whereas Seguam (n = 12) and Atka (n = 12) have a restricted range, at more radiogenic values. There appear to be trends in the Pb isotopic data similar to those seen in $^{87}Sr/^{86}Sr$ between Seguam, Atka, and Adak, with radiogenic values decreasing westward.

Plots of $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ against $^{206}Pb/^{204}Pb$ show rather tight arrays (Myers and Marsh, 1987; Morris and Hart, 1983; Kay and others, 1978). The values of $^{207}Pb/^{204}Pb$ versus $^{206/204}Pb$ partially overlap the normal trend of oceanic ridge basalts and have a steeper slope than the oceanic trend; the Adak values have the greatest amount of scatter. $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ overlaps the MORB array. The total range of values of $^{206}Pb/^{204}Pb$ is also much more restricted (~18.5 to 19.0) than for oceanic island and ridge basalts.

Calc-alkaline and tholeiitic classifications

Coats (1952) studied a small number of samples from Adak and Kanaga and classified them as calc-alkaline. Their alkali-lime (Peacock) index was 63, actually putting them in the "calcic" category (as are most of the subsequently published Aleutian suites). Analyses of these rocks, plotted on an AFM diagram, showed no iron enirchment. Byers (1961) noted an aphyric iron-enriched (12.32 wt. % FeO*) basalt at Okmok, which he compared with tholeiitic basalt of the Columbia River Basalt Group.

Kay (1977) and S. M. Kay and others (1982) applied the FeO*/MgO versus SiO_2 discrimination diagram of Miyashiro (1974) to Aleutian volcanic rocks. Suites having relatively low and constant FeO*/MgO were classified as calc-alkaline, where-

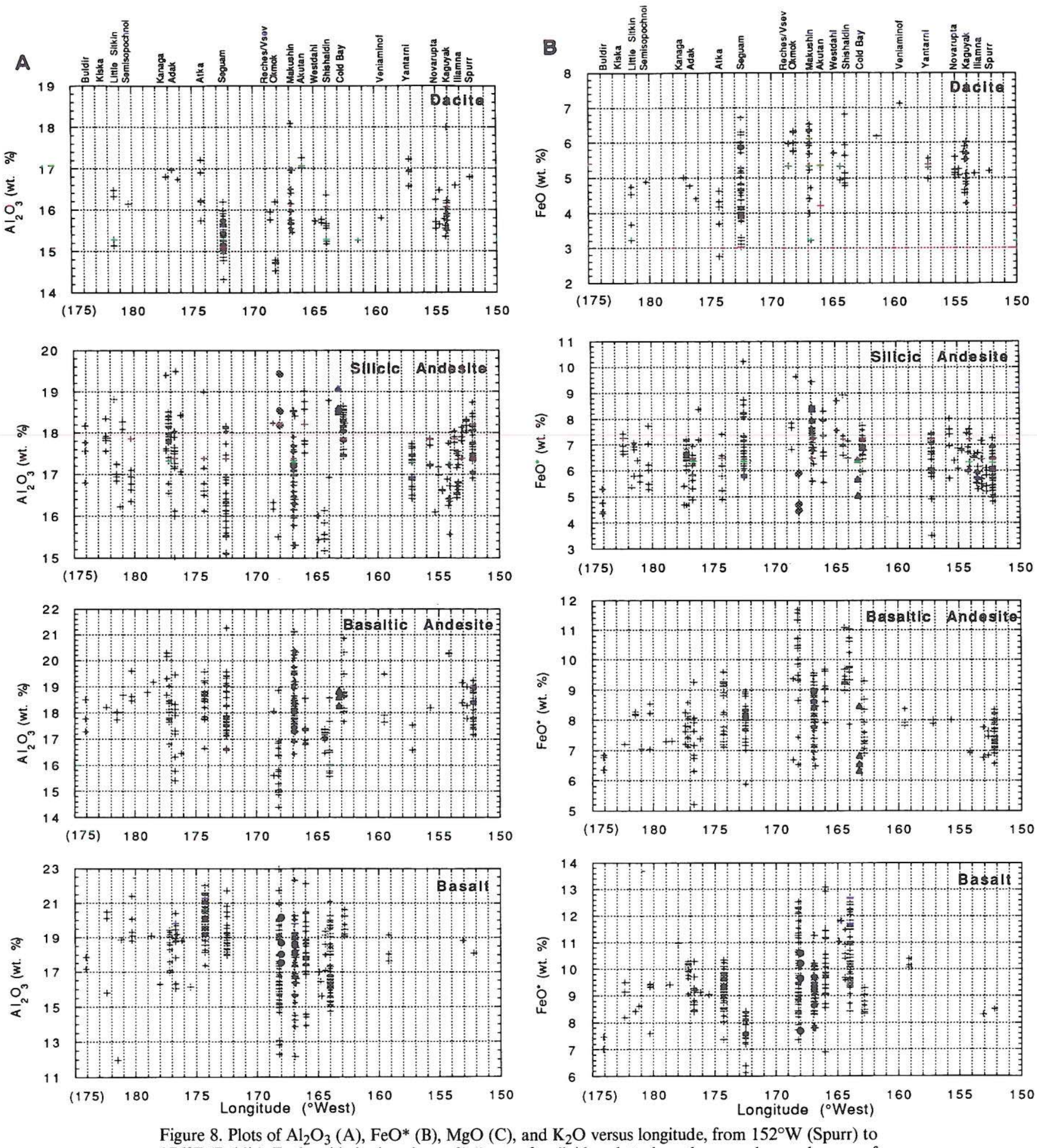
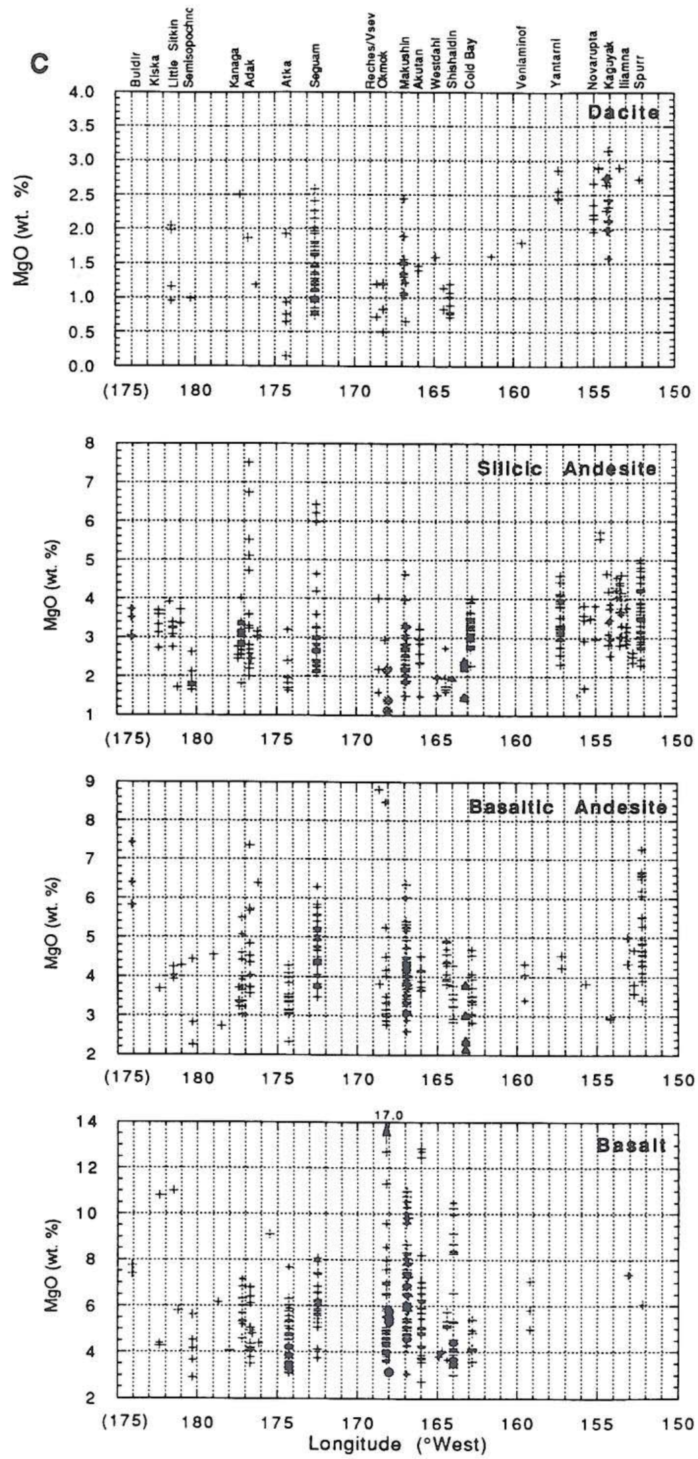
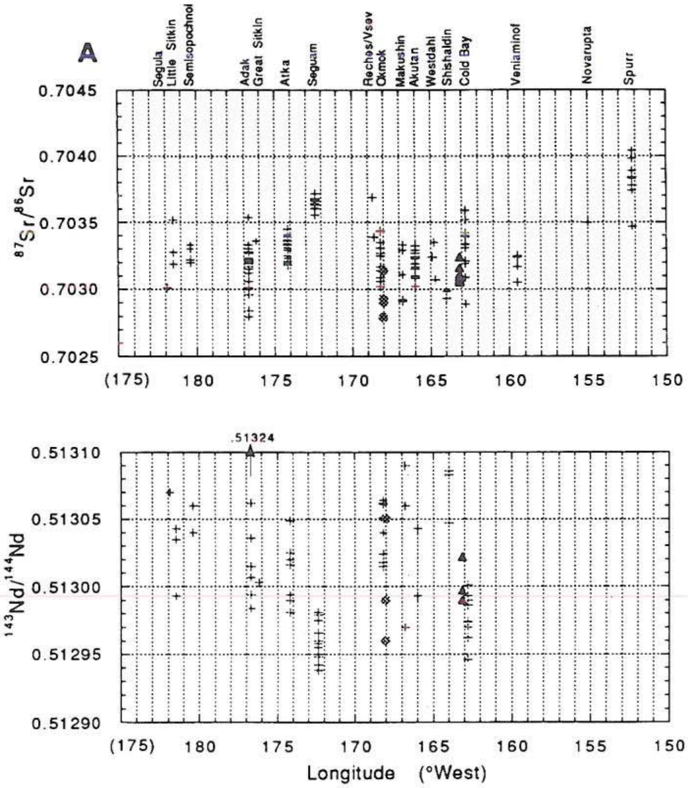


Figure 8. Plots of Al_2O_3 (A), FeO* (B), MgO (C), and K_2O versus longitude, from 152°W (Spurr) to 176°E (Buldir). Each oxide is given in wt. %, "water-free." Also plotted are the second-arc volcanoes of Amak (triangles; 163.15°W) and Bogoslof (solid and hachured circles; 168.03°W). For each oxide, there are separate plots for basalt (<53 wt. % SiO_2), basaltic andesites (53 to 57 wt. % SiO_2), silicic andesites (57 to 63 wt. % SiO_2), and dacites (63 to 70 wt. % SiO_2). Data sources given in Appendix 2.


C
D
Buldir
Kiska
Little Sitkin
Semisopochnoi
Kanaga
Adak
Atka
Seguam
Reches/Vsev
Okmok
Makushin
Akutan
Westdahl
Shishaldin
Cold Bay
Veniaminof
Yantarni
Novarupta
Kaguyak
Iliamna
Spurr
Dacite
Silicic Andesite
Basaltic Andesite
Basalt
MgO (wt. %)
K_2O (wt. %)
17.0
(175) 180 175 170 165 160 155 150
Longitude (°West)



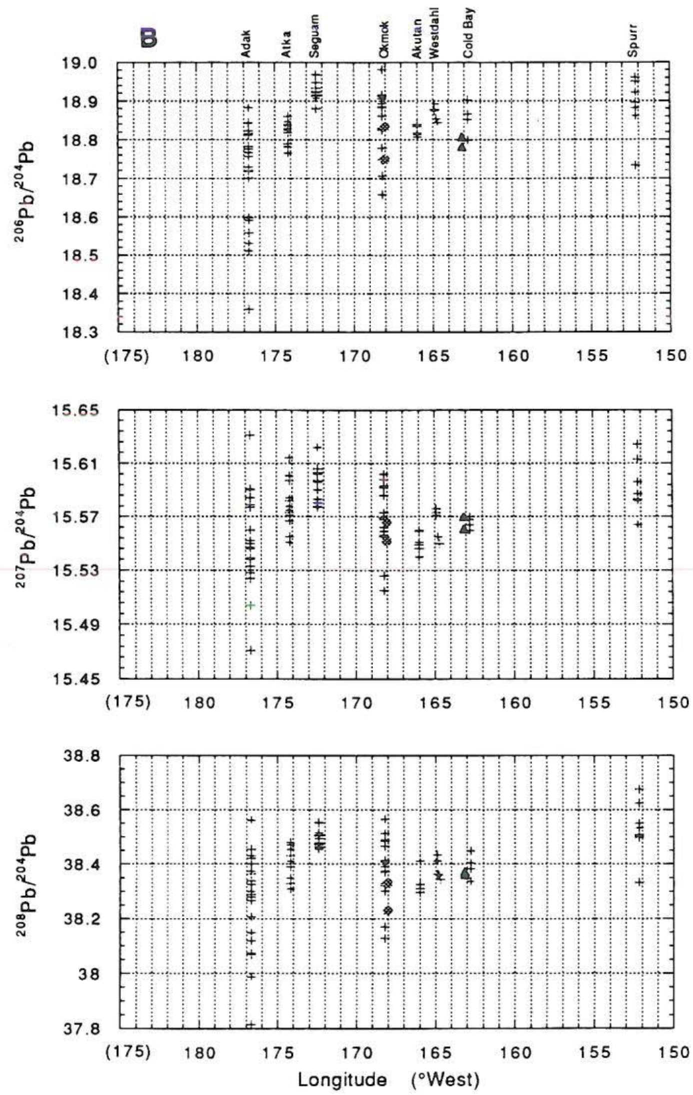


Figure 9. Plot of Sr and Nd (A) and Pb (B) isotopic compositions versus longitude, from 152°W (Spurr) to 176°E (Buldir). Data from Amak (triangles) and Bogoslof (circles) are also plotted. Data sources given in Appendix 2.

as those in which FeO*/MgO increased with silica were termed tholeiites (these rocks differ from MORB tholeiites in their higher K_2O content). S. M. Kay and others (1982) suggested petrographic, major- and trace-element, and volumetric differences among different centers. Large, mainly basaltic volcanic centers are tholeiitic; besides Fe-enrichment, they lack hydrous phenocrysts, have parallel REE patterns, and andesites or dacites are vitrophyric. Calc-alkaline centers are smaller, mainly andesitic, porphyritic with some amphibole, and have nonparallel REE trends. S. M. Kay and Kay (1985) further developed two intermediate classifications: transitional calc-alkaline and transitional tholeiite.

Several volcanic centers (Okmok, Shishaldin, Westdahl/Pogromni, and Seguam) have tholeiitic characteristics. Others such as Cold Bay, Kanaga, and Buldir are calc-alkaline. Myers and others (1985) also found these classifications to be useful for descriptive purposes. Many Aleutian suites, however, do not fit into either category. The dividing line is not clear, particularly in the continental sector of the arc, where Kienle and Swanson (1983) found that FeO*/MgO versus SiO_2 is not a useful discriminator.

Temporal chemical variations

Volcanism on Atka over the last ~7 m.y. shows that each volcanic system began anew with eruption of basalts, followed by andesite and perhaps some dacite. Over this period there has been no change in the fundamental character of the rocks (Marsh, 1980). Just south of Atka, on Amlia Island, there are rocks as old as ~40 Ma, some of which have been studied by Vallier and others (this volume). These rocks show the same diversity of types, but the basalts are virtually identical to those erupted from Korovin (on Atka) over the last 5,000 yrs.

Romick and others (1990) show that, over the past 5 m.y., there has been a decrease in the heterogeneity of Akutan lavas, reflected in bulk composition, mineralogy, and isotopes. The older (1.4 to 4.8 Ma) samples show a wide range in FeO/MgO, particularly in the basalts, plus extreme heterogeneity in plagioclase, olivine, and clinopyroxene compositions. The more recent (~1 Ma to present) volcanic materials show a more restricted range. Although there are less isotopic data for the younger rocks, there appears to be a significant decrease in $^{87}Sr/^{86}Sr$ heterogeneity from the older suite (n = 10, 0.703326 to 0.70302), through the 1-Ma suite (n = 5, 0.703190 to 0.703082), to the recent suite (n = 4, 0.703524 to 0.70342). The modern volcanic rocks show a distinct increase in $^{87}Sr/^{86}Sr$. Lead isotope ($^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$) data is less abundant (five old, two intermediate, one modern), but suggests—from the old to the intermediate stage—decreasing heterogeneity. Romick and others (1990), using incompatible element ratios, suggested that the three temporal groups of lavas were derived from different sources, and that older calc-alkaline or transitional tholeiitic lavas evolved to more depleted (in incompatible elements) and radiogenic (in Sr and Pb) modern tholeiitic lavas.

DSDP ash layers have provided information about Aleutian volcanic episodicity. The submarine ashes analyzed by Scheidegger and others (1980) provide some information, although they represent many combined sources and are likely biased toward more explosive (silicic) compositions. They indicate that volcanism may have become slightly more silicic over the last 1.5 m.y. Over the ~9 m.y. span of these ashes, silicic cycles have occurred every $\sim 10^5$ yr; however, the chemical nature of the volcanism has not changed systematically, and there have been no long-term changes.

Scheidegger and Kulm (1975) and Scheidegger and others (1980) determined chemical compositions for ashes from cores from near Kamchatka (DSDP Site 192), south of Unimak (Site 183), and east of Kodiak Island (Site 178). Because of the prevailing westerly winds, only ashes from the latter two cores probably have any significance for Aleutian volcanism. They conveniently and separately measure the regional volcanism of the oceanic (Site 183) and continental (Site 178) sectors of the arc. These data represent ashes as old as about 9 Ma, whereas the Aleutian volcanics described above are all less than about 3 Ma. Virtually all of the ashes have oxide totals ~96 wt. % (inferred ~4 wt. % H_2O), so some care must be taken when comparing them with subaerial data.

None of the submarine ashes has less than 54 wt. % silica, perhaps due to the lower explosivity of basaltic magma. As in the lavas, the average ash in the more oceanic sector is lower in silica (62 wt. %) than the average ash in the more continental sector (67.6 wt. %). There is significant scatter in Na_2O and CaO, particularly at silica contents above 60 wt. %, which may indicate the effects of alteration by seawater. As in some Aleutian volcanic rocks, there may be two trends in K_2O versus SiO_2.

There are some striking differences, mainly in the mafic (54 to 55 wt. % SiO_2) ashes from Site 183, which have higher TiO_2 and FeO* and lower Al_2O_3 than normally found in Aleutian subaerial volcanics. Ash and lapilli of similar composition were collected in September 1975, 90 km northeast of Shishaldin in the Bering Sea, by a NOAA fisheries research vessel. At that time, Shishaldin was observed to be vigorously erupting. Simkin (written communication, 1976) determined the chemical of this coarse basaltic (50 to 52 wt. % SiO_2) ash as highly enriched in TiO_2 (2.7 to 3.0 wt. %) and FeO* (13.5 to 14.8 wt. %) and depleted in Al_2O_3 (12.4 to 14.7). The apparently anomalous DSDP 183 ashes may actually be geochemical fingerprints of input from the large volcanic centers of Shishaldin and Westdahl on Unimak Island (Site 183 is southeast of Unimak).

PETROLOGY

Phase equilibria

Study of Aleutian lavas and their phenocrysts reveals several petrologic problems that may be resolved by phase equilibria considerations: the common presence of minor amounts of An_{90-100} plagioclase; of olivine either too Fo-rich or too Fo-poor for its host bulk composition; and of diopsidic-salitic clinopyroxene phenocrysts, distinct from normal augitic compositions.

Anorthite-rich plagioclase. Plagioclase dominates the phenocryst assemblage of most Aleutian basalts and andesites, with a wide range of compositions and textures. Cores are richer in An-content than adjacent mantles, although exceptions occur (e.g., Marsh and others, 1990). Drops of An_{15-30} are common at relatively narrow rims. Anorthitic ($>An_{90}$) plagioclase are present in minor amounts in many if not most Aleutian lavas. Such anorthite-rich plagioclase has sometimes been suggested to indicate high water content in arc magmas (e.g., Yoder, 1969). In water-saturated experiments in the simple albite-anorthite system, Yoder (1969) found that water lowers the plagioclase solidus at constant temperature, such that a more anorthitic crystal is produced (i.e., 150 bars H_2O pressure effects a change in the plagioclase from An_{60} to An_{74}).

The effect of water on multicomponent melts of basaltic composition, however, appears to be different. The thermodynamic melt model of Burnham (e.g., 1980) suggests that, for every 10 mole % ($\approx$1 to 1.5 wt. %) increase in water in the melt, the plagioclase becomes only 1 mole % more anorthitic. An unrealistically large amount of water would thus be required to alter crystallization from, say, An_{80} to An_{90}. In fact, high H_2O contents would actually suppress plagioclase crystallization, with clinopyroxene consuming appreciable amounts of Ca; when plagioclase eventually crystallizes, it is of a lower An-content (Crawford and others, 1987). The silicate melt model of Ghiorso (1985) suggests a slight *decrease* in plagioclase An-conent with increasing H_2O. Regardless, both models indicate no major change in An-conent with addition of water to the silicate melt.

In addition, one-atmosphere melting experiments using Aleutian lavas suggest that the soda content of the melt largely dictates plagioclase composition (Marsh and others, 1990). Equilibrium plagioclase compositions of actual high-Al basalts vary, but mainly are An_{75-85}. The distinctive, large, unzoned anorthite crystals require equilibrium with a soda-poor melt, one which has not been identified in the Aleutians.

Olivine. Roeder and Emslie (1970) determined the Fe-Mg distribution between olivine and basaltic liquid and found K_D = 0.30 ± 0.03. The results of oxygen-buffered 1-atmosphere experiments on Aleutian lavas (Peterson and Marsh, unpublished data) similarly indicate K_D = 0.33 (± 0.01). Examination of the Aleutian olivines relative to their host lavas indicates *two* distinct disequilibrium conditions. The presence of olivines too Mg-rich (e.g., Fo_{90-93}) for their host liquid is commonly recognized; however, the second problem is that most olivine phenocrysts found in the Aleutians are generally *too Fe-rich* relative to the host lava (Marsh, unpublished data). Brophy (1984, 1986a) noted this at Cold Bay, where olivines in equilibrium with the host lava should be Fo_{78-80}. The olivines actually preent are Fo_{60-75}, which implies that the actual parental liquid would be more iron-rich (i.e., molar FeO/MgO of 1.1 to 1.8) than the lavas found.

The first disequilibrium condition could be caused by the interaction or contamination of low-Mg basalt by mafic wall rock (i.e., peridotite), which would elevate the bulk Mg content by the addition of Mg-rich olivine or clinopyroxene, with little time for reequilibration. If sufficient time were available, olivine-liquid Fe-Mg reequilibration would eliminate evidence of wall-rock interaction. Keleman (1986, 1990) and Keleman and Ghiorso (1986) have modelled reactions between peridotite and fractionating tholeiitic basalt.

The second disequilibrium condition may result from reequilibration of liquidus olivine with more Fe-rich differentiated residual liquids at higher degrees of crystallinity, such as observed in Hawaiian lava lakes (Moore and Evans, 1967). Alternatively, more Fe-rich magmas may be present at depth but fail to reach the surface.

Magmas having such low Mg-contents do occur at a few Aleutian volcanoes. Fournelle (1988) found that the majority of Shishaldin basalts are Fe-rich, with molar FeO/MgO of 1.1 to 1.4. They contain olivine phenocrysts of Fo_{70-75}, which are at equilibrium with the wallrock composition at the appropriate f_{O_2}. Byers (1961) and Nye and Reid (1986) observed low-Mg basalts at Okmok, and S. M. Kay and others (1982) at Westdahl. Thus, whereas these low-Mg and Fe-enriched basalts are relatively uncommon, they are present at some of the largest Aleutian volcanic centers.

Clinopyroxene. Distinctive green, nonaugitic clinopyroxene occur in lavas from several Aleutian volcanoes (Shishaldin, Okmok, Moffett), as well as volcanoes in Colombia, Nicaragua, and Japan (Fournelle, 1988). They are rich in a diopside or salite component, with elevated levels of Al_2O_3, TiO_2, Cr_2O_3, and Fe^{3+}/Fe^{2+}, and closely resemble those found in the southeastern Alaskan ultramafic layered intrusions. They have been referred to as the salitic trend by Conrad and Kay (1984). At Shishaldin, these clinopyroxene are $Wo_{45-48}En_{42-46}Fs_{8-11}$, with 2.5 to 5.5 wt. % Al_2O_3, 0.5 to 1.4 wt. % TiO_2, and up to 0.7 wt. % Cr_2O_3.

Conrad and Kay (1984) suggested that the presence of this clinopyroxene indicates a high H_2O content in parental Aleutian magmas. Following Irvine's (1974) suggestion that the Duke Island magmas were alkali basalts and thus of low silica activity, Conrad and Kay proposed that this mineral indicated high magmatic water content. This, however, is not the only possible way to cause low silica activity in the magma; a variety of reactions can be written to yield diopside-rich clinopyroxene. There is no need to invoke water, as shown by the following reactions (components in shorthand notation, phases in parentheses):

$$\text{Di (cpx)} + \tfrac{1}{2}\,\text{SiO}_2\text{ (liq)} + \text{Al}_2\text{O}_3\text{ (liq)} = \text{An (pl)} + \tfrac{1}{2}\,\text{Fo (ol)},$$
$$2\text{ Di (cpx)} = \text{CaO (liq)} + 3\text{ SiO}_2\text{ (liq)} + \text{Fo (ol)},$$
$$4\text{ Di (cpx)} + 3\text{ Al}_2\text{O}_3\text{ (liq)} = \text{CaO (liq)} + 3\text{ An (liq)} + 2\text{ Fo (ol)}.$$

Diopsidic clinopyroxene is favored when activities of An and Fo are high, which would be the case if a high-alumina basalt eroded an olivine-rich body.

Silica-undersaturated liquids could be produced by fractionation of a high-Al basalt at pressures greater than a few kilobars (Marsh and others, 1990), for the silica content of fractionating plagioclase may be higher than that of the liquid. Or, a magma reacting with peridotite would be expected to develop a lower silica activity.

Diopsidic-salitic clinopyroxene may also be a reaction product. Manning and Bird (1986), suggested that Lower Zone Skaergaard clinopyroxene ($Wo_{40}En_{47}Fs_{13}$, with low Al_2O_3, TiO_2, and Cr_2O_3) is hydrothermal and formed by reaction of magmatic clinopyroxene with high-temperature aqueous fluids. Fournelle and Marsh (1987) suggested, based upon textural evidence, that Shishaldin diopsidic-salitic clinopyroxenes are products of a low-Mg basalt reacting with Fo-rich olivine bearing Cr-spinel inclusions.

Magmatic intensive parameters

Temperature and water content. Because the exact water content of arc magma is unknown, application of any geothermometer (i.e., plagioclase or olivine) that employs magmatic liquid in its defining reaction cannot be used. The characteristic low titania (<1.1 wt. %) content of arc magmas results in a scarcity of ilmenite, thus limiting the usefulness of the two-oxide (Buddington-Lindsley) geothermometer/oxygen barometer. The two-pyroxene geothermometer is applicable only to andesitic lavas. Direct estimation of magmatic temperatures in basaltic lavas thus is difficult.

The sequence of crystallization in most basalts, especially the stability of plagioclase, is strongly affected by water. At a few kilobars of pressure, saturation with water may cause plagioclase and all mafic phases to crystallize essentially together. A significant amount of water is, however, required for saturation: for a basaltic melt, ~ 7 wt. % at 4 kbar, and ~ 11 wt. % at 8 kbar (Burnham, 1979). The presence of dissolved water in the magma lowers the liquidus temperature to the point that amphibole is stable; this occurs at temperatures below about 1,050°C in a basalt, 950°C in an andesite (Eggler, 1972), and 900°C in a dacite (Merzbacher, 1983; Merzbacher and Eggler, 1984).

In most Aleutian-arc basalts, plagioclase appears on the liquidus well before any other phase. In addition, basaltic plugs containing 60 vol. % crystals show no sign of having entered the stability field of amphibole. These facts suggest that the amount of water in most Aleutian basalts is less than ~ 2 wt. %, and probably less than 1%. Also, as the basalt is undersaturated with water at low pressure, it must be highly undersaturated at high pressures.

Basalts. As described above, pre-eruptive temperatures in basalts are not easy to estimate. Where a pair of pyroxenes or

Fe-Ti oxides occurs, the temperature commonly is in the range 1,100° to 1,200°C (e.g., Brophy, 1986a; Fournelle, 1988). Singer and others (1992b) found temperatures of 1,146° and 1,173°C in a Seguam basalt, using oxygen isotope thermometry on mineral separates. At Okmok, however, Nye and Reid (1986) suggest an unusually high temperature of 1,300° to 1,400°C (at >9 kbar), on the basis of the presumed existence of a (water-free) magma in equilibrium with Fo_{93} olivine. In general, if allowance is made for the crystal-rich nature of the common basalt, extrapolation to a crystal-free state puts the liquidus in the range of 1,200° to 1,250°C.

Andesites. Pre-eruptive temperatures range from about 950 to 1,050°C (Hildreth, 1983; Brophy, 1984, 1987; Fournelle, 1988; Romick and others, 1990; Singer and others, 1992a). Extrapolation to the crystal-free state suggests a liquidus temperature of near 1,200°C; this temperature, plus comparison with Merzbacher and Eggler's (1984) geohygrometer, suggests water contents of 3 to 5 wt. % (Brophy, 1984; Singer and others, 1992a).

Dacites and rhyolites. Crystallizing temperatures in these rocks fall in the range of 800° to 955°C. The dacites occupy the upper half of the range (Anderson, 1975; Hildreth, 1983; Fournelle, 1988; Singer and others, 1992a). Inferred water contents for dacites is 2 to 4 wt. % (Baker and Eggler, 1987; Fournelle, 1988; Singer and others, 1992a). Infrared spectroscopy indicates 2.7 wt. % H_2O in Katmai dacite and 4.5 wt. % HO in rhyolite (Lowenstern, 1990).

Pressure. There is no precise indicator of pressure of crystallization in these lavas. The best procedure is to compare the suite in question with experimentally determined crystallization sequences and coexisting melt compositions (i.e., cotectics) and work out the set of variables (P, H_2O content) most consistent with the rock compositions and mineral assemblages (Marsh, 1976).

The rare occurrence of alumina-rich orthopyroxene (at Cold Bay) suggests that some andesites contain a mineral assemblage that equilibrated at 8 to 10 kbar (Brophy, 1984). Pseudoternary projections suggest that some high-alumina basalts (bulk compositions) reflect a pressure of equilibration (i.e., separation from source) near 20 kbar (Fournelle and Marsh, 1991), whereas the modes and mineral compositions reflect lower pressure (<8 kbar) equilibria (Baker, 1987; Baker and Eggler, 1983, 1987; Gust and Perfit, 1987; Romick and others, 1990). The more silicic lavas appear to have crystallized at lower pressures; Singer and others (1992a), for example, suggest pressures of 3 to 4 kbar for Seguam andesites and 1 to 2 kbar for dacites.

Fugacity of oxygen (f_{O_2}). Oxygen fugacity is commonly calculated from magnetite and ilmenite pairs. Ilmenite is difficult to find in the low-TiO_2 Aleutian volcanic rocks, limiting application of this method for most basalts and andesites. Experimental studies (Sack and others, 1980; Kilinc and others, 1983) have found that the Fe^{3+}/Fe^{2+} of silicic glass is a function of bulk composition, temperature, and f_{O_2}. Oxygen fugacity can be inferred from minimum values of f_{O_2} determined from whole-rock Fe^{3+}/Fe^{2+} composition, assuming that they represent liquid compositions.

A distinctive chemical features of many island-arc lavas is their elevated f_{O_2} relative to ocean-ridge rocks. Whereas ocean-ridge rocks define a clear trend near the quartz-fayalite-magnetite (QFM) buffer, arc lavas show a trend nearly parallel to that buffer, but up to 2 log units higher, above the nickel-nickel oxide (NNO) buffer (Gill, 1981). (In the temperature range of interest, the NNO buffer curve is 0.7 log units above the QFM curve.)

Marsh (1980) suggested, based on the observed phenocryst assemblages, that Atka basalts crystallized about 0.5 to 1 log units above the NNO buffer. Hildreth (1983) found, from FeTi-oxide pairs in 65 of the 1912 Katmai samples, a trend with f_{O_2} above and at a slightly steeper slope than the NNO buffer curve (varying from high to low temperature samples—basalt to rhyolite).

Figure 10a shows the f_{O_2}–temperature relations, determined from FeTi oxides, for Shishaldin, Cold Bay, Okmok, Seguam, and Katmai lavas and pyroclastics.

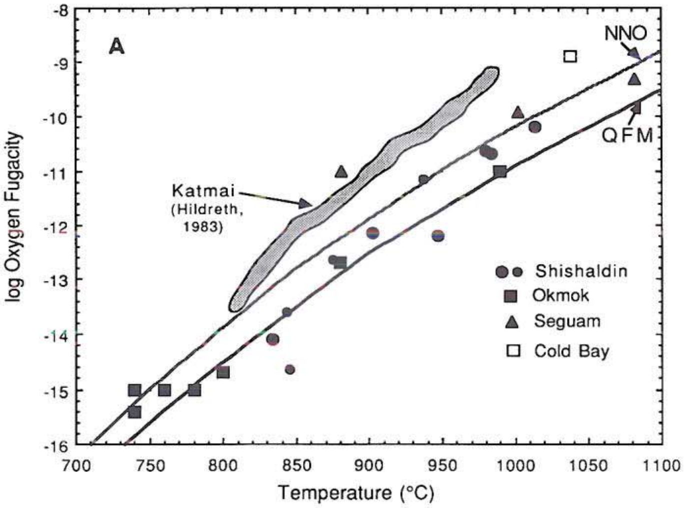
Figure 10b shows the f_{O_2} for Shishaldin lavas calculated from whole-rock Fe^{3+}/Fe^{2+}; it indicates that f_{O_2} lies near the QFM and NNO buffers, consistent with the oxide data. Whereas no FeTi data exist for Atka, whole-rock Fe^{3+}/Fe^{2+} data indicates an f_{O_2} of approximately 1 to 1.5 units above the NNO buffer curve (Fig. 10c).

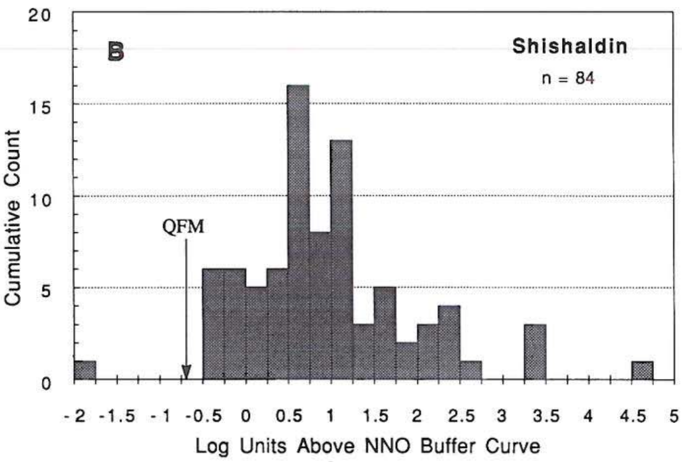
Andesites and dacites at Katmai (Hildreth, 1983), basalts and andesites at Cold Bay (Brophy, 1984, 1986a, 1987), and basalts at Atka (Marsh, 1980, 1982a) crystallized at an f_{O_2} of approximately NNO + 1 log unit, with it decreasing to NNO + 0.1 in rhyolite (Katmai: Hildreth, 1983).

Other Aleutian rocks, however, provide evidence for less oxidized preeruptive conditions. FeTi-oxides in volcanics (primarily basalts) from Shishaldin (Fournelle, 1988), Okmok (Anderson, 1975, written communication, 1987), and Seguam (Singer and others, 1992a) suggest crystallization at an f_{O_2} between the QFM and NNO buffers. It is probably no coincidence that lavas and tephra from these three volcanic centers have the tholeiitic characteristics of iron-enrichment, for the lower f_{O_2} would suppress magnetite (and ilmenite) crystallization.

The presence of sulfides has been documented at one Aleutian volcano, Novarupta. Hildreth (1983) found pyrrhotite (in apparent trace amounts) in mineral separates of the 1912 ejecta; inclusions of pyrrhotite in pyroxenes and oxides were also present. Reconnaissance study also indicates sulfides are present in some Shishaldin basalts (Fournelle, unpublished data).

The apparent absence of sulfides at most Aleutian volcanoes may be due to incomplete studies. However, if the absence is real, it may be a result of magmatic f_{O_2} being in the range of NNO to NNO+1 (or higher) at the particular volcano in question. Carroll and Rutherford (1987, 1988) found that magmatic sulfur speciation changes drastically, from sulfide to sulfate, over this f_{O_2} range. At higher f_{O_2}, anhydrite replaces pyrrhotite. Anhydrite has not been found yet in Aleutian volcanic rocks, but it is unstable in a wet environment. It has been found elsewhere, in freshly collected pumices (El Chichon: Luhr and others, 1984)





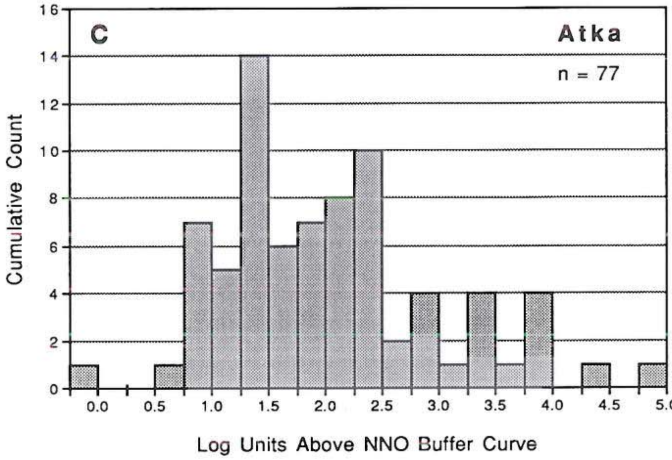


Figure 10A, Plot of log oxygen fugacity versus temperature (°C) determined from coexisting FeTi oxides from five Aleutian volcanoes. Data are from Cold Bay (Brophy, 1984), Shishaldin (Fournelle, 1988), Katmai (Hildreth, 1983), Okmok (Anderson, 1975, written communications, 1987), and Seguam (Singer and others, 1991). For Shishaldin, the larger symbol indicates grains >100 microns; smaller symbol, <100 microns. Also shown are the nickel-nickel oxide (NNO) and quartz-fayalite-magnetite (QFM) buffer curves. B, Calculated oxygen fugacities of Shishaldin volcanic rocks based upon whole-rock Fe^{3+}/Fe^{2+} following the experimental algorithm of Sack and others (1980) and Kilinc and others (1983). These values are consistent with f_{O_2} values between the QFM and NNO buffers, determined from ilmenite-magnetite pairs. (Three data above NNO+5 not included.) C, Calculated oxygen fugacities of Atka volcanic rocks, based upon whole-rock Fe^{3+}/Fe^{2+}, as for B. These values are consistent with modal mineralogy of Atka volcanic rocks, which suggests equilibrium 1 to 2 log units above the nickel-nickel oxide (NNO) buffer.

and in inclusions within pumice phenocrysts (Nevado de Ruiz: Fournelle, 1990a).

Like isotopic composition, f_{O_2} is inherited by a magma from its source rock. It is difficult to change f_{O_2}, except along the usual pseudobuffer or indicator curves by changing temperature, although it could be done by assimilating large amounts of highly reduced or oxidized wall rock, or by reaction of variable oxidation state sulfur phases in the magma (e.g., Whitney and Stormer, 1983).

High-alumina basalts and plagioclase accumulation

Many Aleutian high-alumina basalts (i.e., 17 to 22 wt. % Al_2O_3) contain 25 to 50 vol. % plagioclase. Crawford and other's (1987) global study concluded that such high-Al basalts are cumulates and they calculated that the groundmass of Aleutian high-Al basalt should be andesitic-dacitic. Later studies on Aleutian lavas have accepted these conclusions (Brophy, 1989a). Brophy (1989b) developed a model to explain plagioclase retention in crustal magma chambers.

In a study of plagioclase-rich recent lavas from Shishaldin, Fournelle and Marsh (1991) found no evidence of plagioclase accumulation. Groundmass and mineral separates were studied. The groundmass separates were basaltic, not andesitic or dacitic. Shishaldin high-Al basalts have positive Eu-anomalies, but they do not match those expected by plagioclase addition, and may instead be a signature of clinopyroxene in an eclogitic source (Brophy, 1986b). There is no evidence of disequilibrium between plagioclase and liquid/groundmass in the high-Al basalts. Plots of Al_2O_3, CaO, and Na_2O versus modal plagioclase do not correspond with plagioclase addition to high-Mg basalt or dacite. On the other hand, several older (~3 to 5 Ma) high-Al basalts from an adjacent glaciated unit show some evidence for plagioclase accumulation in a high-Mg basalt; Fournelle and Marsh (1991) suggested that this was a possible occurrence in the early development of the volcanic center.

Xenoliths and xenocrysts

True ultramafic (peridotite) xenoliths are uncommon in island arcs. Swanson and others (1987) examined the characteristics of Aleutian xenoliths from Kanaga and Adak. Those at Kanaga are in Tertiary rocks and not related to current volcanism; at Adak, ultramafic inclusions are present in two brecciated units (not lavas). Gabbroic xenoliths from Adak have been studied by Conrad and others (1983) and Conrad and Kay (1984), who suggested they were cumulate phases of early arc magmas.

More common, however, are individual crystals that stand out as foreign to their host lava.

Olivine. These olivine crystals are highly magnesian (Fo_{90-93}), large (0.1 to 0.2 cm), sometimes slightly strained and have a more fayalitic rim. Somewhat less magnesian (Fo_{85}), smaller, and highly corroded olivine crystals also sometimes occur in andesitic lavas that otherwise lack olivine (e.g., Adak: Marsh, 1976; Cold Bay: Brophy, 1984). These olivines are not in equilibrium with their host rocks, and appear to be from mantle peridotite or oceanic crust.

Based upon a global study of olivine minor element composition, Simkin and Smith (1970) suggested that xenocrysts had less than 0.10 wt. % Ca (elemental). Nye and Reid (1986) used this criterion to interpret Okmok Fo_{92-93} as nonxenocrystic. However, in a more recent study of xenocrystic olivine, Bodinier and others (1987) found 0.15 to 0.16 wt. % Ca in olivine cores in some garnet and spinel lherzolites. We suggest that the olivine Ca-content limit of 0.10 wt. % should not be used to preclude Fo-rich olivine from being characterized as xenocrystic.

Plagioclase. Plagioclase more calcic than An_{90}, and certainly above An_{95}, is xenocrystic, and could be from the underlying arc crust or from fragments of older decoupled slabs. Megacrysts of An_{90} have been found in a basalt dredged from the flank of a seamount just south of the Aleutian Trench in the Adak fracture zone (Fournelle, unpublished data).

Quartz. The presence of rounded ~1 cm quartz in high-Mg Shishaldin basalt indicates contamination (Fournelle, 1988).

Also, Marsh (unpublished data) found a Si-rich glassy bomb at Adak (sample AD97), with 88.6 wt. % SiO_2, 3.1 wt. % Al_2O_3, and 5.2 wt. % H_2O+CO_2, and having Sr and Pb isotopic compositions within the range of all Adak samples.

PETROGENESIS

Introduction

Discussions of arc petrogenesis have focused either on the deep source at the slab-mantle wedge interface, or upper level (crustal) processes. Three key source components have been hypothesized: the subducted oceanic crust, subducted sediments, and the overlying mantle wedge (Fig. 11). Upper level processes (e.g., fractionation, magma mixing, assimilation) are considered possible modifiers of the primary magma compositions. Presumably, by inverting the process and sorting out the upper level effects, the nature of the original source can be deciphered. This approach, however, assumes that the long ~100-km ascent through the lithosphere (i.e., upper mantle) will have a *negligible* effect on the composition of ascending magma. This assumption would be true only if the magma and the lithosphere were identical in composition and at the same temperature, which is unlikely, as then there would be no buoyancy to force the magma to the surface. Instead, it appears most reasonable that chemical and thermal interactions between compositionally distinct magma and mantle may significantly alter the primary magma as it develops its ascent conduit (Marsh, 1978; Myers and others, 1985; Singer and others, 1989).

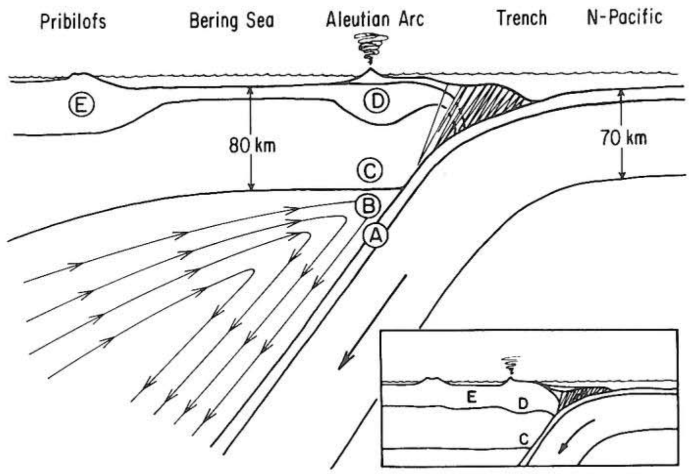


Figure 11. Simplified cross section through the oceanic part of the Aleutian arc. Possible regions of magma production and interaction are: A, subducted oceanic crust (with layer of sediments); B, asthenosphere ("mantle wedge"); C, arc lithosphere; D, subvolcanic front welt; E, "continental" crust. The insert shows a cross section through the Alaska Peninsula. Heavy arrow indicates motion of subducted plate. Streamlines indicate convective motion of the asthenosphere as it is coupled with downward motion of the subducted slab. (From von Drach and others, 1986.)

Diapir transport

How does magma travel upward from its source? There is no sign of earthquakes leading upward through the lithosphere (as at Hawaii) from near the subducting plate, even beneath the smallest, youngest volcanic centers. This suggests a passive mode of magma transfer such as diapirism. Enough heat must be carried upward to soften the wall sufficiently to allow the body to travel fast enough to reach the surface while still partially molten (Marsh, 1978, 1982b). To allow this, at least several bodies must successively travel essentially the same path through the lithosphere. As the initial mush begins ascent it undergoes increased melting as it tries to rise further away from its solidus. Once melting progresses to about 50 vol. %, the remaining solids can be repacked to free a magmatic liquid. The possibility that viscous compaction of the solids releases a liquid at very small amounts of melting seems much less likely (Brophy and Marsh, 1986).

The sharpness of the volcanic front and the fairly regular spacing of the volcanic centers imply that the magma may be transported to the surface via gravitational instability. The magmatic source "layer" produces a series of (ideally) evenly spaced diapirs, whose spacing and size is controlled by the thickness and viscosity of the layer and by the viscosity of the overlying material. An experimental and analytical study of this process, as applied to the Aleutians (Marsh, 1979b), suggests that the magmatic zone is thin ($<\sim500$ m), narrow (~10 to 30 km), highly viscous ($\sim10^{12}$ to 10^{14} poise), and produces diapirs of 3 to 5 km in radius. Marsh (1979b) calculated the local mantle viscosity as $\sim4 \times 10^{20}$ poise, which is in close accord with studies using glacial rebound.

Singer and others (1989) calculated that 3- to 6-km-radius diapirs, periodically ascending at 4×10^{-8} to 2×10^{-7} m/s, could establish magmatic conduits from 120 km depth to the crust in 60,000 to 900,000 years. In the process, large volumes of magma must solidify in the lithospheric mantle.

Evolution of volcanic plumbing systems

The rigors of fully penetrating the lithosphere almost guarantees that the first eruptions of a new volcanic center will be of magma that has interacted strongly, chemically and physically, with its wall rock (Marsh and Leitz, 1979). The first extrusives would be expected to have a heterogeneous isotopic, trace element, and bulk chemical signature, the net result of the original source material plus any lithospheric or asthenospheric wall rock it has encountered during ascent, as well as enhanced crystallization due to the initial cool state of the conduit walls. Bogoslof, appearing in 1796, is an excellent example.

Each magma traveling this general pathway will also react with the wall rock, but each time to a lesser degree as the conduit walls themselves are "contaminated" by solidified magma (Marsh and Kantha, 1978). As the conduit is gradually heated, successive diapirs may experience different crystallization sequences, which may in turn yield different liquid lines of descent (Singer and others, 1989). Passage of enough magma will produce a thermally and chemically insulated conduit, thus allowing magma to erupt that could carry the geochemical signature of its source (Fig. 12). Small volcanic centers, then, are likely to have "dirty" lithospheric plumbing systems, whereas larger, mature centers should have relatively "clean" systems (Myers and others, 1985).

An example of the foregoing model is Adak (Marsh, 1982a), a small to medium-sized volcanic center (~30 km^3) in the central Aleutians. Strontium and lead isotopic data, along with major- and trace-element data, show a good deal of scatter. One of the lowest exposed units—a breccia—that issued from Mount Moffett has an unusually low $^{87}Sr/^{86}Sr$ value of 0.7028; i.e., sample ADK-53 of Kay (1978), is andesitic (55 wt. % SiO_2), but contains high (5.58 wt. %) MgO, 150 ppm Ni, anomalously high Sr (1,783 ppm), and is dominated by large, unzoned fragments of Mg-rich clinopyroxene. Slightly higher in this same section is an andesitic flow containing abundant Fo-rich olivines (Marsh, 1976; sample AD-48); these features are consistent with strong interaction with lithospheric wall rocks.

A larger center, such as Atka (~200 km^3), on the other

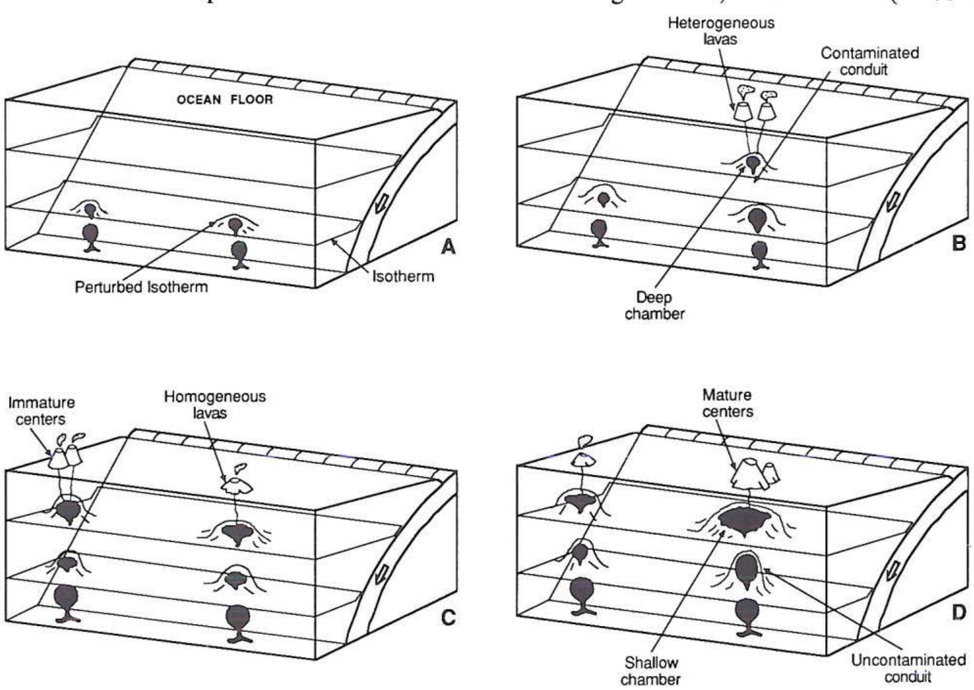


Figure 12. Schematic representation of the evolution of an Aleutian volcanic center. A, early stages of magma ascent with significant thermal and chemical interaction between magma and wall rock; B, eruption of the first highly contaminated lavas from deep-seated magma bodies; C, magma bodies are closer to the surface and erupted lavas are less contaminated; D, fully developed plumbing system. Eruptions are from well-established shallow magma chambers. Material in the plumbing system itself has been removed or chemically modified. Lavas erupted at this stage are petrographically and chemically homogeneous. (After Myers and others, 1985.)

hand shows a tight, restricted range of Sr and Pb isotopic data and smooth trends of major and trace elements. No xenoliths or xenocrysts (besides An-rich plagioclase) have been found in these lavas.

Volcanic centers that become too big (e.g., Okmok, 200 to 300 km^3; and Shishaldin, 300 to 400 km^3), may have so much magma fluxing the uppermost lithosphere that they erode and entrain fragments of ultramafic or mafic wall rock. Those lavas would show a distinct chemical signature that is indistinguishable from a peridotite source (e.g., Fo-rich olivine with chromite inclusions; high whole-rock Mg-content, Ni and Cr contents)—unless the magma were erupted so quickly that the intact wall rock (xenoliths or xenocrysts) were carried to the surface. This appears to have happened at Shishaldin, where chromite and clinopyroxene compositions in high-Mg basalts resemble those of the southeastern Alaskan zoned ultramafic complexes and of the Border Ranges ultramafic and mafic complex of Burns (1985). Based upon this evidence and geophysical studies of the Border Ranges fault (Fisher, 1981; Marlow and Cooper, 1983), Fournelle (1988) suggested that the Border Ranges ultramafic and mafic complex may be present below Shishdalin volcano. A similar scenario was suggested by Arculus and others (1983) at Mount Lamington (Papua New Guinea), where Mg-rich xenoliths were found in andesitic lavas. The Papuan Ultramafic Belt crops out within 10 km of the volcano, and geophysical evidence indicates it lies 12 to 16 km below the volcano.

The source of Aleutian magmas

The controversy over the source of Aleutian magmas is essentially the same today as when it was posed by Coats in 1962: whether the subducted slab yields an aqueous fluid (containing silica, Sr, Rb, Ba, and Pb) that causes melting of the mantle wedge; or whether the slab itself melts, yielding a silicate fluid (melt), containing some water and Sr, Rb, Ba, and Pb. This problem has been addressed using isotopes, trace elements, rare-earth elements, major elements, and magmatic intensive parameters, but seldom have all been considered simultaneously. McCulloch and Perfit (1981, p. 176) summed up the situation: "Although the ostensibly simple model of producing the Aleutian arc magmas by directly melting of a source consisting of partially altered MORB and several percent entapped sediment satisfies the Nd, Sr, and Pb isotopic constraints, it must also be compatible with the major and trace element data." The two main hypotheses are discussed below.

Slab dewatering–peridotite melting model. This hypothesis holds that an aqueous fluid from the subducted slab causes melting of the overlying mantle wedge, yielding a high-Mg basaltic melt that ascends to near (or above) the crust-mantle boundary, where fractionation can produce either high-Al basalts, or combined assimilation/fractional crystallization yields calc-alkaline andesites (Perfit and others, 1980a, 1980b; Gill, 1981; Nye and Reid, 1986; Gust and Perfit, 1987; Kay and Kay, this volume). Key to this model are the occasional arc basalts having elevated MgO (>10 wt. %), Ni (>100 ppm), and Cr (>100 ppm) and bearing $Fo_{90\text{-}92}$ olivine. These unusual lavas are cited as evidence of an abundant but commonly untapped parental magma.

The presence of Fo-rich olivine in high-Mg arc basalts, however, is not conclusive evidence that these basalts were derived by partial melting of mantle peridotite (Myers, 1988). Such high-Mg basalts can be produced by low-Mg basalt interacting with ultramafic material in the lithosphere below the volcano (Arculus and others, 1983; Fournelle, 1988).

A major question of arc petrogenesis is how to melt the cold subducted slab. Mantle wedge metasomatism has been suggested (e.g., Tatsumi and others, 1986), which assumes that sufficient water is still present (having not been dehydrated earlier), and that the released aqueous phase travels vertically upward. Oxburgh and Turcotte (1976) suggest instead that at shallow depths, water may migrate downwards, deeper into the slab. Only at much greater depths (up to 700 km), could it be released upwards. Eggler (1989), on the other hand, suggests that any shallow (<75 km) slab-derived fluids will travel back up the slab, whereas fluids released at greater depths will be swept downward with the induced mantle flow.

If water-induced melting of peridotite is the process responsible for Aleutian arc magmatism, and if the commonly observed Aleutian high-alumina basalts are derived by fractionation from the high-magnesian basalt in equilibrium with this peridotite, then two consequences should be evident. First, the water content of the high-alumina basalts (the products of fractionation) should be unusually high if water-saturated melting of peridotite is invoked. If the peridotite melt is undersaturated with respect to water, then there should be noticeably high water contents of the high-alumina basalt. The common phenocryst assemblage in Aleutian basalts—lack of amphibole and abundance of plagioclase—suggests however that there is less than 1-2 wt. % H_2O in the basaltic magma at several kilobars of pressure, and therefore a parental magma must have much less than 0.5-1 wt. % H_2O at high pressures. Second, if water induces melting of peridotite by lowering the solidus temperature, then the temperatures recorded in the derivative high-alumina basalts should be even lower. Oxide, pyroxene, and oxygen isotope geothermometry show, however, that basalt temperatures are in the range 1,100° - 1,200°C (Brophy, 1986a; Fournelle, 1988; Singer and others, 1991, 1991b). The temperature of the parental magma should then be greater than 1,200°C.

A common belief is that the oxidized nature of arc lavas is a result of water fluxing the mantle. Studies of hydrothermal alteration of oceanic crust at mid-ocean ridge vents, coupled with the role of arc volcanoes in the global sulfur cycle (e.g., recent SO_2-rich eruptions of El Chichon, Nevado del Ruiz, and Mt. Pinatubo volcanoes), however, suggest that some iron in the oceanic crust may have been oxidized by the reduction of anhydrite (precipitated from seawater) to sulfide (Albarede and Michard, 1989) prior to subduction.

Kay and Kay (1989; this volume) have developed a model of asthenospheric circulation, in which the mantle flow above the slab is opposite that of all geophysical models (e.g., Hsui and others, 1983; Tatsumi and others, 1986; see Fig. 11). Their model also is not consistent with the geophysical constraints of back-arc spreading, which occurs behind some arcs but not behind the Aleutian arc where the crust is too old, cold, and thick.

Eggler (1989) presented evidence that partial melts of the mantle wedge (with $H_2O + CO_2$), immediately above the slab at 75 to 125 km, are silica-poor and highly alkalic—carbonatitic, alkali carbonatitic, or melilitic. Increased melting would yield more olivine-rich melts approaching alkali picrites. These melts have little in common with arc-lava compositions.

Slab melting–lithospheric mining model. This model, which we prefer, holds that an upper portion of the subducted slab (oceanic crust plus sediment, metamorphosed to quartz eclogite) melts, and that high-alumina basalts represent primary magma compositions (Marsh, 1982a; Marsh and Carmichael, 1974; Myers, 1988; Johnston, 1986; Brophy and Marsh, 1986 [their Table 6 has a summary of the pros and cons of the peridotite versus quartz eclogite source models]).

Myers and others (1985) and Myers and Marsh (1987) suggested that slab-derived magmas may be compositionally modified by interaction with peridotite during ascent through the lithosphere (Fig. 12), producing high-Mg basalt. Assimilation of mafic material by less mafic magma may occur, as has been described by Arculus and others (1983), Kudo (1983), Kelemen and Sonnenfeld (1983), Evans (1985), Kelemen (1986), and Kelemen and Ghiorso (1986). The lack of olivine on (or near) the liquidus of Aleutian high-Al basalt suggests that it was never in equilibrium with peridotite.

Most Aleutian arc lavas display enrichment in alkalis with unfractionated REE. Brophy and Marsh (1986) explained the unfractionated (La/Yb) REE of Aleutian high-Al basalts lavas using a model of diapiric ascent and extraction of melt outside the garnet stability field. They calculated that the Rb and Ba levels (and most major elements) of the basalts were consistent with 40 to 60% melting of altered MORB plus 5 % Pacific pelagic sediment, although modelled Sr abundances are below those observed.

Hsui and others (1983) modelled the thermal regime above the cold subducting slab and suggested that enough hot mantle material could come in contact with the slab to begin to melt its upper surface at a depth of 100 to 150 km.

The crystal-rich nature of Aleutian basalts has generated some skepticism about the existence of primary high-Al basaltic liquids (e.g., Crawford and others, 1987). The study by Fournelle and Marsh (1991) of recent Shishaldin basalts suggests that these high-Al basalts are not plagioclase cumulates, and could be primary to ~20 kbar. On the other hand, older (~3 to 5 Ma) high-Al bsalts from an erosional remnant near Shishaldin have some features consistent with the addition of plagioclase to a high-Mg basalt. These features are consistent with the developmental history of the lithospheric conduit feeding the volcanic center; earlier magmas may have been more Mg-rich, owing to lithospheric interaction. It is possible that high-Al basalts developed from these magmas by fractionation of olivine and clinopyroxene, as in the model of Gust and Perfit (1987), with or without addition of plagioclase.

Production of island-arc andesite

Island arcs were once thought to consist dominantly of andesites. The Aleutian island arc, however, is dominated by high-Al basalts through andesites and only the Alaska Peninsula–Cook Inlet arc section is dominated by andesitic compositions (Fig. 6). Isotopically, the andesites (and dacites) are identical to the basalts, and could be derived from basalts by fractionation. Baker and Eggler (1987), however, found that at high pressure (8 kbar), Atka andesites and dacites can not be produced by fractionation of basalt owing to thermal divides (i.e., separation of minerals drives the residual liquid to lower silica content).

Gill (1981) concluded that calc-alkaline arc andesites are produced by a complex series of mechanisms: (1) deep interaction of slab-derived water or melt with the mantle wedge: melting, separation, and ascent; (2) intermediate level fractionation during ascent: stagnation at base of the crust or within it; (3) upper level fractionation. This general model has been applied to Aleutian andesites by Kay and others (1982). Brophy (1990) further developed it in relation to Kanaga volcanism by proposing that: A, low-Al (~high-Mg) basalt replenishes a magma chamber at 2 to 5 kbar and high-Al basalt is formed by fractionation; and B, significant amounts of crustal assimilation produce dacitic liquids. Eruption at this point, with conduit mixing of high-Al basalt and dacite, yields mixed andesites. This open-system model accounts for the eruption only of hybrid calc-alkaline basaltic andesites and andesites (54 to 63 wt. % SiO_2) and the absence of basalts or dacites.

There is another explanation for the abundance of calc-alkaline andesites, requiring no mixing of unseen end members. Kudo (1983) suggested that the textural evidence required to support magma mixing in Japanese calc-alkaline andesites could be explained adequately by assimilation processes. Mafic gabbroic inclusions, used as evidence that the magma has undergone fractionation, instead may be a result of the magma cooling as it attempted to assimilate contaminating xenoliths. This approach has gained further support by the thermodynamic modelling of Kelemen and Ghiorso (1986), who showed that combined fractional crystallization and assimilation of mafic wall rock (i.e., Fo_{90}) by a dioritic liquid with 2 wt. % H_2O could produce the calc-alkaline AFM trend. Kelemen (1990) suggested this as a viable mechanism for producing calc-alkaline suites at convergent margins worldwide.

Singer and others (1991) suggest a less complex, closed system fractionation model to explain Seguam tholeiitic andesites and dacites. Small batches of basalt or basaltic andesite separate from a larger ~3 to 4 kbar reservoir; by cooling, they can evolve to more silicic compositions, without crustal assimilation or mixing. Integral here is the extended nature of the crust at Seguam, compared to the unextended crust at Kanaga and Adak (Singer and Myers, 1990).

FUTURE WORK

Although the Aleutian island arc is one of the best studied oceanic arcs in the world, many parts of it have not been mapped or extensively sampled. For example, the recent volcanic products on Tanaga, the largest island in the western Aleutians, have been described only briefly (Coats and Marsh, 1984). In addition, a large segment of the central arc, from Amuktu to Kagamil, is virtually unknown geologically. Recent field mapping just to the west of this segment has shown that the volcanic centers of the central portion of the arc may exhibit geologic and geochemical characteristics unlike the rest of the arc (Myers and Singer, 1987; Singer and Myers, 1988). Owing to the ruggedness of the islands, their remoteness, the hardships of field work, and the short field seasons, geologic investigations to date have been mostly reconnaissance in nature. Consequently, even the "mapped" islands may hold surprises for future field geologists.

APPENDIX 1. SUMMARY OF THE NUMBERS OF CHEMICAL ANALYSES, BY INDIVIDUAL VOLCANIC CENTER (WEST OF 161°W), USED IN THIS CHAPTER

Volcanic Center	Major	Trace Elements						Isotopes		
	Elements	Rb	Sr	Ba	Cr	Ni	REE	Sr	Nd	Pb
Cold Bay	48	29	29	21	9	8	1	13	8	4
Shishaldin	88	88	88	88	88	88	23	3	3	0
Fisher Caldera	24	24	24	24	24	24	0	0	0	0
Westdahl, Pogromni	6	5	5	2	0	0	2	3	0	5
Akutan	46	36	37	18	18	17	14	19	3	7
Makushin	177	106	93	23	17	18	4	6	3	0
Okmok	72	57	57	24	19	46	12	12	8	15
Recheshnoi	18	1	13	13	13	12	1	1	0	0
Vsevidof	4	2	2	2	0	0	2	1	0	0
Seguam	181	171	181	177	177	85	25	15	15	19
Atka	84	25	35	34	34	34	20	24	7	12
Kasatochi	1	0	0	0	0	0	0	0	0	0
Great Sitkin	5	1	1	0	0	0	0	1	1	0
Adak	47	31	32	30	24	26	29	23	13	21
Kanaga	52	44	44	44	33	1	34	0	0	0
Bobrof	7	0	0	0	0	0	0	0	0	0
Semisopochnoi	15	14	14	5	12	14	5	5	4	0
Little Sitkin	13	13	13	3	0	13	3	3	3	0
Segula	2	2	2	1	0	2	1	1	1	0
Kiska	7	5	5	0	0	5	0	0	0	0
Buldir	9	0	0	0	0	0	0	0	0	0
Bogoslof	9	5	5	5	2	2	3	4	3	2
Amak	7	7	7	4	4	1	0	7	3	2

Note: Specific trace elements and isotopes are indicated to show the relative proportions of chemical data between various volcanic centers.

APPENDIX 2. SOURCES OF GEOCHEMICAL DATA FOR ALEUTIAN VOLCANIC CENTERS

Spurr	Nye and Turner, 1990
Redoubt	Forbes and others, 1969; Kienle and others, 1983
Iliamna	Kienle and others, 1983
Augustine	Kienle and others, 1983
Douglas	Kienle and others, 1983
Fourpeaked	Kienle and others, 1983
Kaguyak	Kienle and others, 1983
Devils Desk	Kienle and others, 1983
Kukak	Kienle and others, 1983
Denison	Kienle and others, 1983
Snowy	Kienle and others, 1983
Novarupta	Hildreth, 1983, 1987
Trident	Forbes and others, 1969
Martin	Kienle and others, 1983
Kejulik	Kienle and others, 1983
Yantarni	Riehle and others, 1987
Veniaminof	Kay and others, 1982; Yount and others, 1985
Pavlof	Anderson, 1975
Cold Bay	Marsh, 1976; Kay, 1977, Kay and others, 1978; Marsh and Leitz, 1979; Kay and others, 1982; Brophy, 1984, 1986a, 1987
Shishaldin	Fournelle, 1988; Fournelle and Marsh, 1991
Fisher Caldera	Fournelle, 1990b, unpublished data; T. Miller, personal comm., 1985
Westdahl, Pogromni	Kay, 1977; Kay and others, 1978
Akutan	McCulloch and Perfit, 1981; Romick, 1982; Romick and others, 1990
Makushin	Drewes and others, 1961; DeLong, 1974; Perfit, 1977; Perfit and others, 1980a; McCulloch and Perfit, 1981; Nye and others, 1986; Gust and Perfit, 1987
Okmok	Byers, 1959, 1961; Kay, 1977; Kay and others, 1978; McCulloch and Perfit, 1981; Nye, 1983; Kay and Kay, 1985; Nye and Reid, 1986, 1987; T. Miller, personal comm., 1985
Recheshnoi	Byers, 1959, 1961; Franks, 1981; Kay, 1977; Kay and others, 1978
Vsevidof	Byers, 1959, 1961; Kay, 1977; Kay and others, 1978
Seguam	Myers, unpublished data; Singer and others, 1992a
Atka	Marsh, 1980, 1982a, unpublished data; Myers and others, 1985, 1986; von Drach and others, 1986; Baker and Eggler, 1983, 1987; Myers and Marsh, 1987; Myers and Frost, unpublished data
Kasatochi	Kay and Kay, 1985
Great Sitkin	Simons and Mathewson, 1953; Marsh, 1976; Kay and Kay, 1985; von Drach and others, 1986
Adak	Coats, 1952; Marsh, 1976; Kay, 1977, 1978; Kay and others, 1978; Kay and others, 1982; Conrad and Kay, 1984; Kay and Kay, 1985; Myers and others, 1985; Kay and others, 1986; von Drach and others, 1986; Debari and others, 1987; Myers and Marsh, 1987; Myers and Frost, unpublished data
Kanaga	Coats, 1952; Fraser and Barrett, 1959; DeLong, 1974; Kay, 1977; Brophy, 1989a, 1990, unpublished data
Bobrof	Kay and others, 1982
Tanaga	Coats and Marsh, 1984
Semisopochnoi	Coats, 1959b; DeLong, 1974; DeLong and others, 1985
Little Sitkin	Snyder, 1959; DeLong, 1974; White and Patchett, 1984
Segula	Nelson, 1959; DeLong, 1974; McCulloch and Perfit, 1981
Kiska	Coats and others, 1961; DeLong, 1974
Buldir	Coats, 1953; Kay and Kay, 1985
Bogoslof	Fenner, 1926; Arculus and others, 1977; Kay, 1977; Kay and others, 1978; Marsh and Leitz, 1979; McCulloch and Perfit, 1981; von Drach and others, 1986
Amak	Marsh and Leitz, 1979; Morris and Hart, 1983; von Drach and others, 1986
submarine	Scholl and others, 1976

REFERENCES CITED

Albarede, F., and Michard, A., 1989, Hydrothermal alteration of the oceanic crust, *in* Hart, S. R., and Gulen, L., eds. Crust/mantle recycling at convergence zones: Dordrecht, Kluwer Academic, p. 29–36.

Anderson, A. T., 1975, Some basaltic and andesites gases: Reviews of Geophysics and Space Physics, v. 13, p. 37–55.

Aramaki, S., and Ui, T., 1983, Alkali mapping of the Japanese Quaternary volcanic rocks: Journal of Volcanology and Geothermal Research, v. 18, p. 549–560.

Arculus, R. J., DeLong, S. E., Kay, R. W., Brooks, C., and Sun, S. S., 1977, The alkalic rock suite of Bogoslof Island, eastern Aleutian arc, Alaska: Journal of Geology, v. 85, p. 177–186.

Arculus, R. J., Johnson, R. W., Chappell, B. W., McKee, C. O., and Sakai, H., 1983, Ophiolite-contaminated andesites, trachybasalts, and cognate inclusions of Mount Lamington, Papua New Guinea; Anhydrite-amphibole-bearing lavas and the 1951 cumulodome: Journal of Volcanology and Geothermal Research, v. 18, p. 215–247.

Baker, D. R., 1987, Depths and water content of magma chambers in the Aleutian and Mariana island arcs: Geology, v. 15, p. 496–499.

Baker, D. R. and Eggler, D. H., 1983, Fractionation paths of Atka (Aleutian) high-alumina basalts; Constraints from phase relations: Journal of Volcanology and Geothermal Research, v. 18, p. 387–404.

—— , 1987, Compositions of anhydrous and hydrous melts coexisting with plagioclase, augite, and olivine or low-Ca pyroxene from 1 atm to 8 kbar; Application to the Aleutian volcanic center of Atka: American Mineralogist, v. 72, p. 12–28.

Becker, G. F., 1898, Reconnaissance of the gold fields of southern Alaska, with some notes on general geology: U.S. Geological Survey 18th Annual Report, v. 3.

Black, L. T., 1981, Volcanism as a factor in human ecology; The Aleutian case: Ethnohistory, v. 28, p. 313–340.

Boudinier, J. -L., Dupuy, C., Dostal, J., and Meriet, C., 1987, Distribution of trace transition elements in olivine and pyroxenes from ultramafic xenoliths; Application of microprobe analysis: American Mineralogist, v. 72, p. 902–913.

Brophy, J. G., 1984, The chemistry and physics of Aleutian arc volcanism; The Cold Bay volcanic center, southwestern Alaska [Ph.D. thesis]: The Johns Hopkins University, 422 p.

—— , 1986a, The Cold Bay volcanic center, Aleutian volcanic arc; I, Implications for the origin of hi-alumina arc basalt: Contributions to Mineralogy and Petrology, v. 93, p. 368–380.

—— , 1986b, Source rock clinopyroxene, Eu anomalies, and the origin of hi-alumina basalt: EOS, v. 67, p. 405.

—— , 1987, The Cold Bay Volcanic Center, Aleutian volcanic arc; II, Implications for fractionation and mixing mechanism in calc-alkaline andesite genesis: Contributions to Mineralogy and Petrology, v. 97, p. 378–388.

—— , 1989a, Can high alumina basalt be derived from low alumina arc basalt? Evidence from Kanaga Island, Aleutian arc, Alaska: Geology, v. 17, p. 333–336.

—— , 1989b, Basalt convection and plagioclase retention, a model for the generation of high-alumina arc basalt: Journal of Geology, v. 97, p. 319–329.

—— , 1990, Andesites from northeastern Kanaga Island, Aleutians; Implications for calc-alkaline fractionation mechanisms and magma chamber development: Contributions to Mineralogy and Petrology, v. 104, p. 568–581.

Brophy, J. G., and Marsh, B. D., 1986, On the origin of high-alumina arc basalt and the mechanics of melt extraction: Journal of Petrology, v. 27, p. 763–789.

Burk, C. A., 1965, Geology of the Alaska Peninsula; Part 1, Island arc and continental margin: Geological Society of America Memoir 99, 250 p.

Burnham, C. W., 1979, The importance of volatile constituents, *in* Yoder, H. S., ed., The evolution of the igneous rocks: Princeton University Press, p. 439–482.

—— , 1980, The nature of multicomponent silica melts, *in,* Richard, D. T. and Wickman, F. E., eds., Chemistry and geochemistry of solutions at high temperatures and pressures: Pergamon Press, Physics and Chemistry of the Earth, v. 13 and 14, p. 197–229.

Byers, F. M., Jr., 1959, Geology of Umnak and Bogoslof Islands, Aleutian Islands, Alaska: U.S. Geological Survey Bulletin 1028-L, p. 267–369.

—— , 1961, Petrology of three volcanic suites, Umnak and Bogoslof Islands, Aleutian Islands, Alaska: Geological Society of America Bulletin, v. 72, p. 93–128.

Carroll, M. R., and Rutherford, M. J., 1987, The experimental stability of igneous anhydrite; Experimental results and implications for sulfur behavior in the 1982 El Chichon trachyandesite and other evolved magmas: Journal of Petrology, v. 28, p. 781–801.

—— , 1988, Sulfur speciation in hydrous experimental glasses of varying oxidation state; Results from measured wavelength shifts of sulfur X-rays: American Mineralogist, v. 73, p. 845–849.

Coats, R. R., 1947, Reconnaissance geology of some western Aleutian Islands: U.S. Geological Survey Alaskan Volcano Investigations, p. 95–105.

—— , 1950, Volcanic activity in the Aleutian arc: U.S. Geological Survey Bulletin 974-B, p. 35-49.

—— , 1952, Magmatic differentiation in Tertiary and Quaternary volcanic rocks from Adak and Kanaga Islands, Aleutian Islands, Alaska: Geological Society of America Bulletin, v. 63, p. 485–514.

—— , 1953, Geology of Buldir Island, Aleutian Islands, Alaska: U.S. Geological Survey Bulletin 989-A, p. 1–25.

—— , 1956, Geology of northern Adak Island, Alaska: U.S. Geological Survey Bulletin 1028-C, p. 45–67.

—— , 1959, Geologic reconnaissance of Semisopochnoi Island, western Aleutian Islands, Alaska: Geological Survey Bulletin 1028-O, p. 477–519.

—— , 1962, Magma type and crustal structure in the Aleutian arc, *in* The crust of the Pacific basin: American Geophysical Union Monograph 6, p. 92–109.

Coats, R. R., and Marsh, B. D., 1984, Reconnaissance geology and petrology of northern Tanaga, Aleutian Islands, Alaska: Geological Society of America Abstracts with Programs, v. 16, p. 474.

Coats, R. R., Nelson, W. H., Lewis, R. Q., and Powers, H. A., 1961, Geologic reconnaissance of Kiska Island, Aleutian Islands, Alaska: U.S. Geological Survey Bulletin 1028-R, p. 563–581.

Conrad, W. K., and Kay, R. W., 1984, Ultramafic and mafic inclusions from Adak Island; Crystallization history, and implications for the nature of primary magmas and crustal evolution of the Aleutian arc: Journal of Petrology, v. 25, p. 88–125.

Conrad, W. K., Kay, S. M., and Kay, R. W., 1983, Magma mixing in the Aleutian arc; Evidence from cognate inclusions and composite xenoliths: Journal of Volcanology and Geothermal Research, v. 8, p. 279–295.

Crawford, A. J., Falloon, T. J., and Eggins, S., 1987, The origin of island arc high-alumina basalts: Contributions to Mineralogy and Petrology, v. 97, p. 417–430.

Curtis, G. H., 1968, The stratigraphy of the ejecta from the 1912 eruption of Mount Katmai and Novarupta, Alaska, *in* Coats, R. R., Hay, R. L., and Anderson, C. A., eds., Studies in volcanology: Geological Society of America Memoir 166, p. 153–210.

Dall, W. H., 1870, Alaska and its resources: London, Lee and Shepard, 627 p.

—— , 1918, Reminiscences of Alaska volcanoes: Scientific Monthly, v. 7, p. 80–90.

DeBari, S., Kay, S. M., and Kay, R. W., 1987, Ultramafic xenoliths from Adagdak Volcano, Adak, Aleutian Islands, Alaska; Deformed igneous cumulates from the Moho of an island arc: Journal of Geology, v. 95, p. 329–341.

DeLong, S. E., 1974, Distribution of Rb, Sr and Ni in igneous rocks, central and western Aleutian Islands, Alaska: Geochimica et Cosmochimica Acta, v. 38, p. 245–266.

DeLong, S. E., Perfit, M. R., McCulloch, M. T., and Ach, J., 1985, Magmatic, evolution of Semisopochnoi Island, Alaska; Trace-element and isotopic constraints: Journal of Geology, v. 93, p. 609–618.

Drewes, H., Fraser, G. D., Synder, G. L., and Barnett, H. F., Jr., 1961, Geology of Unalaska Island and adjacent insular shelf, Aleutian Islands, Alaska: U.S. Geological Survey Bulletin 1028-S, p. 583–763.

Dumond, D. E., 1979, People and pumice on the Alaska Peninsula, *in* Sheets, P. D., and Grayson, D. K., eds., Volcanic activity and human ecology: New York, Academic Press, p. 373–392.

Eggler, D. H., 1972, Water-saturated and undersaturated melting relations in a Paricutin andesite and an estimate of water content in the natural magma: Contributions to Mineralogy and Petrology, v. 34, p. 261–171.

——, 1989, Influence of H_2O and CO_2 on melt and fluid chemistry in subduction zones, *in* Hart, S. R., and Gulen, L., eds., Crust/mantle recycling at convergence zones: Dordrecht, Kluwer Academic, p. 97–104.

Engdahl, E. R., 1973, Relocation of intermediate depth earthquakes in the central Aleutians by seismic ray tracing: Nature, v. 245, p. 23–25.

Evans, C. A., 1985, Magmatic "metasomatism" in peridotites from the Zambales ophiolite: Geology, v. 13, p. 166–169.

Fenner, C. N., 1923, The origin and mode of emplacement of the great tuff deposit of the Valley of Ten Thousand Smokes: National Geographic Society, Contributed Technical Paper, Katmai Series, no. 1, 74 p.

——, 1926, The Katmai magmatic province: Journal of Geology, v. 34, p. 673–772.

Finch, R. H., 1934, Shishaldin Volcano, Proceedings of the Fifth Pacific Science Conference, Canada: Toronto, University of Toronto Press, v. 3, p. 2369–2376.

——, 1935, Akutan Volcano: Zeitschrift fur Vulkanologie, v. 16, p. 155–160.

Fisher, M. A., 1981, Location of the Border Ranges fault southwest of Kodiak Island, Alaska: Geological Society of America Bulletin, v. 92, p. 19–30.

Forbes, R. B., Ray, D. K., Katsura, T., Matsumoto, H., Haramura, H., and Furst, M. J., 1969, The comparative chemical compositions of continental vs. island arc andesites in Alaska, *in* McBirney, A. R., ed., Proceedings of the Andesite Conference: Oregon State Department of Geology and Mineral Industries, Bulletin 65, p. 111–120.

Fournelle, J. H., 1988, The geology and petrology of Shishaldin Volcano, Unimak Island, Aleutian arc, Alaska [Ph.D. thesis]: The Johns Hopkins University, 529 p.

——, 1990a, Anhydrite in Nevado del Ruiz November 1985 pumice; Relevance to the sulfur problem: Journal of Volcanology and Geothermal Research, v. 42, p. 189–201.

——, 1990b, Geology and geochemistry of Fisher Caldera, Unimak Island, Aleutians; Initial results: EOS, v. 71, p. 1,698.

Fournelle, J. H. and Marsh, B. D., 1987, Diopsidic cpx produced by assimilation of Fo-rich olivine by magmas at Shishaldin Volcano: Implications for Alaskan layered complexes: EOS Transactions of the American Geophysical Union, v. 68, p. 1525.

——, 1991, Shishaldin Volcano; Aleutian high-alumina basalts and the plagioclase accumulation question: Geology, v. 19, p. 234–237.

Franks, A. L., 1981, A petrological study of Mt. Recheschnoi Volcano, Umnak Island, Alaska [M.S. thesis]: New York, Columbia University, 35 p.

Fraser, G. D., and Barnett, H. D., 1959, Geology of the Delarof and westernmost Andreanof Islands, Aleutian Islands, Alaska: U.S. Geologic Survey Survey Bulletin 1028-I, p. 211–245.

Fujita, K., Engdahl, E. R., and Sleep, N. H., 1981, Subduction zone calibration and teleseismic relocation of thrust zone events in the central Aleutian Islands: Bulletin of the Seismological Society of America, v. 71, p. 1805–1825.

Geist, E. L., Childs, J. R., and Scholl, D. W., 1987, The evolution and petroleum geology of the Amlia and Amukta intra-arc basins: Marine Petroleum Geology, v. 4, p. 334–352.

——, 1988, The origin of summit basins of the Aleutian ridge; Implications for block rotation of an arc massif: Tectonophysics, v. 7, p. 327–341.

Gerlach, T. M., Westrich, H. R., and Casadevall, T. J., 1990, High sulfur and chlorine magma during the 1989–90 eruption of Redoubt Volcano, Alaska: EOS Transactions of the American Geophysical Union, v. 43, p. 1702.

Ghiorso, M. S., 1985, Chemical mass transfer in magmatic processes; I, Thermodynamic relations and numerical algorithms: Contributions to Mineralogy and Petrology, v. 90, p. 107–120.

Gill, J. B., 1981, Orogenic andesites and plate tectonics: Springer-Verlag, 390 p.

Grewingk, C., 1850, Beitrag zur Kenntniss der orographischen und geognostishen Beschaffenheit der nordwest-Kuste Amerikas mit den anliegenden Inseln, St. Petersburg: reprint from Mineralogisches Gesellschaft St. Petersburg Verhandlungen, 1848–49, 351 p.

Griggs, R. F., 1922, The Valley of Ten Thousand Smokes: Washington, D.C., National Geographic Society, 340 p.

Gust, D. A., and Perfit, M. R., 1987, Phase relations of a high-Mg basalt from the Aleutian island arc; Implications for primary island arc basalts and high-Al basalts: Contributions to Mineralogy and Petrology, v. 97, p. 7–18.

Hein, J. R., Scholl, D. W., and Miller, J., 1978, Episodes of Aleutian ridge explosive volcanism: Science, v. 199, p. 137–141.

Hein, J. R., McLean, H., and Vallier, T., 1984, Reconnaissance geology of Southern Atka Island, Aleutian Islands, Alaska: U.S. Geological Survey Bulletin 1609, 19 p.

Hildreth, W., 1983, The compositionally zoned eruption of 1912 in the Valley of Ten Thousand Smokes, Katmai National Park, Alaska: Journal of Volcanology and Geothermal Research, v. 18, p. 1–56.

——, 1987, New perspectives on the eruption of 1912 in the Valley of Ten Thousand Smokes, Katmai National Park, Alaska: Bulletin of Volcanology, v. 49, p. 680–693.

Howell, D. G., Jones, D. L., and Schermer, E. R., 1985, Tectonostratigraphic terranes of the circum-Pacific region, *in* Howell, D. G., ed., Tectonostratigraphic terranes of the circum-Pacific region: Houston, Circum-Pacific Council for Energy and Mineral Resources, p. 3–30.

Hsui, A. T., Marsh, B. D., and Toksoz, M. N., 1983, On melting of the subducted oceanic crust; Effects of subduction induced mantle flow: Tectonophysics, v. 99, p. 207–220.

Hubbard, B., 1932a, Alaska's Shishaldin; The first ascent of "Smoky Moses": Saturday Evening Post, August 13, p. 8+.

——, 1932b, Mush, you malemutes!: New York, The American Press, 179 p.

——, 1936, Cradle of the storms: London, George G. Harrap and Co., 285 p.

Irvine, T. N., 1974, Petrology of the Duke Island ultramafic complex, southeastern Alaska: Geological Society of America Memoir, v. 138, 240 p.

Jacob, K. H., Nakamura, K., and Davies, J. N., 1977, Trench-volcano gap along the Alaska-Aleutian arc, facts and speculations on the role of terrigenous sediments for subduction, *in* Talwani, M., and Pitman, W. C., III, eds., Island arcs, deep sea trenches and back arc basins: American Geophysical Union, p. 243–258.

Jaggar, T. A., Jr., 1908, Journal of the technology expedition to the Aleutian Islands: The Technology Review, v. 10, p. 1–37.

——, 1932, Aleutian eruptions, 1930–1932: Volcano Letter, no. 375, p. 1–4.

Johnston, A. D., 1986, Anhydrous P-T phase relations of nearly-primary high-alumina basalt from the South Sandwich Islands: Contributions to Mineralogy and Petrology, v. 92, p. 368–382.

Kay, R. W., 1977, Geochemical constraints on the origin of Aleutian magma, *in* Talwani, M., and Pitman, W. C., III, eds., Island arcs, deep sea trenches and back-arc basins: Washington, D.C., American Geophysical Union, p. 229–242.

——, 1978, Aleutian magnesian andesites; Melts from subducted Pacific Ocean crust: Journal of Volcanology and Geothermal Research, v. 4, p. 117–132.

Kay, R. W., and Kay, S. M., 1989, Recycled continental crustal components in Aleutian arc magmas, implications for crustal growth and mantle heterogeneity, *in* Hart, S. R., and Gulen, L., eds., Crust/mantle recycling at convergence zones: Dordrecht, Kluwer Academic, p. 145–161.

Kay, R. W., Sun, S. S., and Lee-Hu, C. -N., 1978, Pb and Sr isotopes in volcanic rocks from the Aleutian Islands and Pribilof Islands, Alaska, Geochimica et Cosmochimica Acta, v. 42, p. 263–274.

Kay, R. W., Rubenstone, J. L., and Kay, S. M., 1986, Aleutian terranes from Nd isotopes: Nature, v. 322, p. 605–609.

Kay, S. M., and Kay, R. W., 1985, Aleutian tholeiitic and calc-alkaline magma series; I, The mafic phenocrysts: Contributions to Mineralogy and Petrology, v. 90, p. 276–290.

Kay, S. M., Kay, R. W., and Citron, G. P., 1982, Tectonic controls on tholeiitic and calc-alkalic magmatism in the Aleutian arc: Journal of Geophysical Research, v. 87, p. 4051–4072.

Kelemen, P. B., 1986, Assimilation of ultramafic rocks in subduction-related magmatic arcs: Journal of Geology, v. 94, p. 829–843.

——, 1990, Reaction between ultramafic rock and fractionating basaltic magma; I, Phase relations, the origin of calc-alkaline magma series, and the formation of discordant dunite: Journal of Petrology, v. 31, p. 51–98.

Kelemen, P. B., and Ghiorso, M. S., 1986, Assimilation of peridotite in zoned calc-alkaline plutonic complexes; Evidence from the Big Jim complex, Washington Cascades: Contributions to Mineralogy and Petrology, v. 94, p. 12–28.

Kelemen, P. B., and Sonnfeld, M. D., 1983, Stratigraphy, structure, petrology and local tectonics, central Ladakh, NW Himalaya: Schweizer Mineralogische Petrographische Mitteilungen, v. 63, p. 267–287.

Kennedy, G. C., and Waldron, H. H., 1955, Geology of Pavlof Volcano and vicinity, Alaska: U.S. Geological Survey Bulletin 1028-A, p. 1–19.

Kienle, J., and Swanson, S. E., 1983, Volcanism in the eastern Aleutian arc; Late Quaternary and Holocene centers, tectonic setting and petrology: Journal of Volcanology and Geothermal Research, v. 17, p. 393–432.

Kienle, J., Swanson, S. E., and Pulpan, H., 1983, Magmatism and subduction in the eastern Aleutian arc, *in* Shimozuru, D., and Yokoyama, I., eds., Arc volcanism; Physics and tectonics: Tokyo, Terra Scientific Publishing Company, p. 191–224.

Kilinc, A., Carmichael, I.S.E., Rivers, M. L., and Sack, R. O., 1983, Ferric-ferrous ratio of natural silicate liquids equilibrated in air: Contributions to Mineralogy and Petrology, v. 83, p. 136–140.

Kudo, A. M., 1983, Origin of Calc-alkalic andesites, Nasu zone, northeastern Japan; Kuno revisited: Geochemical Journal, v. 17, p. 51–62.

Lowenstern, J. B., 1990, Pre-eruptive water content of high-silica rhyolite and dacite from the 1912 eruption at the Valley of Ten Thousand Smokes, Alaska: EOS Transactions of the American Geophysical Union, v. 71, p. 1690.

Maddren, A. G., 1919, Sulphur on Unalaska and Akun Islands and near Stepovak Bay, Alaska: U.S. Geological Survey Bulletin 692-E, p. 268–298.

Manning, C. E., and Bird, D. K., 1986, Hydrothermal clinopyroxenes of the Skaergaard intrusion: Contributions to Mineralogy and Petrology, v. 92, p. 437–447.

Marlow, M. S., and Cooper, A. K., 1983, Wandering terranes in southern Alaska; The Aleutia microplate and implications for the Bering Sea: Journal of Geophysical Research, v. 88, p. 3439–3446.

Marsh, B. D., 1976, Some Aleutian andesites, their nature and source: Journal of Geology, v. 84, p. 27–45.

——, 1978, On the cooling of ascending andesitic magma: Philosophical Transaction Royal Society of London A, v. 288, p. 611–625.

——, 1979a, Island arc volcanism: American Scientist, v. 67, p. 161.

——, 1979b, Island arc development; Some observations, experiments, and speculations: Journal of Geology, v. 87, p. 687–713.

——, 1980, Geology and petrology of northern Atka, Aleutian Islands, Alaska: Geological Society of America, Abstracts with Programs, v. 12, p. 476.

——, 1982a, The Aleutians, *in* Thorpe, R. S., ed., Orogenic andesites: John Wiley & Sons, p. 99–114.

——, 1982b, On the mechanics of igneous diapirism, stoping, and zone melting: American Journal of Science, v. 282, p. 808–855.

——, 1987, Petrology and evolution of the N.E. Pacific including the Aleutians: Pacific Rim Congress 87, p. 309–315.

Marsh, B. D., and Carmichael, I.S.E., 1974, Benioff zone magmatism: Journal of Geophysical Research, v. 79, p. 1196–1206.

Marsh, B. D., and Kantha, L. H., 1978, On the heat and mass transfer from an ascending magma: Earth and Planetary Science Letters, v. 39, p. 435–443.

Marsh. B. D., and Leitz, R. E., 1979, Geology of Amak Island, Aleutian Islands, Alaska: Journal of Geology, v. 87, p. 715–723.

Marsh, B. D., Fournelle, J. H., Myers, J. D., and Chou, I. -M., 1990, On plagioclase thermometry in island-arc rocks; Experiments and theory, *in* Spencer, R. J., and Chou, I. -M., eds., Fluid-mineral interactions, A tribute to H. P. Eugster: Geochemical Society Special Publication, v. 2, p. 65–83.

McClelland, L., Simkin, T., Summers, M., Nielsen, E., and Stein, T. C., 1989, Global Volcanism, 1975–1985: Englewood Cliffs, New Jersey, Prentice Hall, 655 p.

McCulloch, M. T., and Perfit, M. R., 1981, $^{143}Nd/^{144}Nd$, $^{87}Sr/^{86}Sr$ and trace element constraints on the petrogenesis of Aleutian island arc magmas: Earth and Planetary Science Letters, v. 56, p. 167–179.

McNutt, S. R., and Jacob, K. H., 1986, Determination of large-scale velocity structure of the crust and upper mantle in the vicinity of Pavlof Volcano, Alaska: Journal of Geophysical Research, v. 91, p. 5013–5022.

Merrill, G. P., 1885, On hornblende andesites from the new volcano of Bogoslof Island in Bering Sea: Proceedings of U.S. National Museum, v. 8 (1886), p. 31–33.

Merzbacher, C., 1983, Water-saturated and -undersaturated phase relations of the Mount St. Helens dacite magma erupted on May 18, 1980, and an estimate of the pre-eruptive water content [M.S. thesis]: University Park, Pennsylvania State University, 54 p.

Merzbacher, C., and Eggler, D. H., 1984, A magmatic geohygrometer; Application to Mount St. Helens and other dacitic magmas: Geology, v. 12, p. 587–590.

Miller, T. P., and Smith, R. L., 1984, Calderas of the eastern Aleutian arc: Transactions of the American Geophysical Union, v. 64, no. 45, p. 87.

Minster, J. B., Jordan, T. H., Molnar, P., and Haines, E., 1974, Numerical modelling of instantaneous plate tectonics: Journal of Royal Astronomical Society, v. 36, p. 541–576.

Miyashiro, A., 1974, Volcanic rock series in island arcs and active continental margins: American Journal of Science, v. 274, p. 321–355.

Moore, J. G., and Evans, B. W., 1967, The role of olivine in the crystallization of the prehistoric Makaopuhi theoleiite lava lake, Hawaii: Contributions to Mineralogy and Petrology, v. 15, p. 202–223.

Morris, J. D., and Hart, S. R., 1983, Isotopic and incompatible element constraints on the genesis of island arc volcanics from Cold Bay and Amak Island, Aleutians, and implications for mantle structure: Geochimica et Cosmochimica Acta, v. 47, p. 2015–2030.

Myers, J. D., 1988, Possible petrogenetic relations between low- and high-MgO Aleutian basalts: Bulletin of Geological Society of America, v. 100, p. 1040–1053.

Myers, J. D., and Frost, C. D., 1989, Trace and rare earth element constraints on the origin and evolution of Aleutian arc magmas: EOS Transactions of the American Geophysical Union, v. 17, p. 721.

Myers, J. D., and Marsh, B. D., 1987, Aleutian lead isotopic data; Additional evidence for the evolution of lithospheric plumbing systems: Geochimica et Cosmochimica Acta, v. 51, p. 1833–1842.

Myers, J. D., and Singer, B. S., 1987, Seguam Island, central Aleutian Islands, Alaska; I, Geologic field relations: EOS Transactions of the American Geophysical Union, v. 68, p. 1525.

Myers, J. D., Marsh, B. D., and Sinha, A. K., 1985, Strontium isotopic and selected trace element variations between two Aleutian volcanic centers (Adak and Atka); Implications for the development of arc volcanic plumbing systems: Contributions to Mineralogy and Petrology, v. 91, p. 221–234.

——, 1986, Geochemical and strontium isotopic characteristics of parental Aleutian arc magmas; Evidence from the basaltic lavas of Atka: Contributions to Mineralogy and Petrology, v. 94, p. 1–11.

Nakamura, K., 1977, Volcanoes as possible indicators of tectonic stress orientation; Principle and proposal: Journal of Volcanology and Geothermal Research, v. 2, p. 1–16.

Nakamura, K., Plafker, G., Jacob, K. H., and Davies, J. N., 1980, A tectonic stress trajectory map of Alaska using information from volcanoes and faults: Bulletin of the Earthquake Research Institute, v. 55, p. 89–100.

Neal, R. J., and Swanson, S. W., 1983, Petrology and geochemistry of Westdahl and Pogromni Volcanoes, Unimak Island, Alaska: EOS Transactions of the American Geophysical Union, v. 64, p. 893.

Nelson, W. H., 1959, Geology of Segula, Davidof and Khvostof Islands, Alaska: U.S. Geological Survey Bulletin 1028-K, p. 257–266.

Nye, C. J., 1983, Petrology and geochemistry of Okmok and Wrangell Volcanoes, Alaska [Ph.D. thesis]: University of California at Santa Cruz, 208 p.
Nye, C. J., and Reid, M. R., 1986, Geochemistry of primary and least fractionated lavas from Okmok Volcano, central Aleutians; Implications for magmagenesis: Journal of Geophysical Research, v. 91, p. 10271–10287.
—— , 1987, Corrections to "Geochemistry of primary and least fractionated lavas from Okmok Volcano, central Aleutians; Implications for magmagenesis": Journal of Geophysical Research, v. 92, p. 8182.
Nye, C. J., and Turner, D. L., 1990, Petrology, geochemistry, and age of the Spurr volcanic complex, eastern Aleutian arc: Bulletin of Volcanology, v. 52, p. 205–226.
Nye, C. J., Swanson, S. E., and Reeder, J. W., 1986, Petrology and geochemistry of Quaternary volcanic rocks from Makushin Volcano, central Aleutian arc: Alaska Division of Geological and Geophysical Surveys, Open-File Report 86-80, p. 1–106.
Oxburgh, E. R., and Turcotte, D. L., 1976, The physico-chemical behavior of the descending lithosphere: Tectonophysics, v. 32, p. 107–128.
Perfit, M. R., 1977, The petrochemistry of igenous rocks from the Cayman Trench and the Captains Bay pluton, Unalaska Island; Their relation to tectonic processes at plate margins [Ph.D. thesis]: New York, Columbia University, 273 p.
Perfit, M. R., Brueckner, H., Lawrence, J. R., and Kay, R. W., 1980a, Trace element and isotopic variations in a zoned pluton and associated volcanic rocks, Unalaska Island, Alaska; A model for fractionation in the Aleutian calc-alkaline suite: Contributions to Mineralogy and Petrology, v. 73, p. 69–87.
Perfit, M. R., Gust, D. A., Bence, A. E., Arculus, R. J. and Taylor, S. R., 1980b, Chemical characteristics of island arc basalts; Implications for mantle sources: Chemical Geology, v. 30, p. 227–256.
Reinink-Smith, L. M., 1990a, Relative frequency of Neogene volcanic events as recorded in coal partings from the Kenai lowland, Alaska; A comparison with deep-sea core data: Geological Society of America Bulletin, v. 102, p. 830–840.
—— , 1990b, Mineral assemblages of volcanic and detrital partings in Tertiary coal beds, Kenai Peninsula, Alaska: Clays and Clay Minerals, v. 38, p. 97–108.
Riehle, J. R., Yount, M. E., and Miller, T. P., 1987, Petrography, chemistry, and geologic history of Yantarni Volcano, Aleutian volcanic arc, Alaska: U.S. Geological Survey Bulletin 1761, 27 p.
Roeder, R. L., and Emslie, E. G., 1970, Olivine-liquid equilibrium: Contributions to Mineralogy and Petrology, v. 29, p. 275–289.
Romick, J. D., 1982, The igneous petrology and geochemistry of northern Akutan Island, Alaska [M.S. thesis]: Fairbanks, University of Alaska, 151 p.
Romick, J. D., Perfit, M. R., Swanson, S. E., and Shuster, R. D., 1990, Magmatism in the eastern Aleutian arc; Temporal characteristics of igneous activity on Akutan Island: Contributions to Mineralogy and Petrology, v. 104, p. 700–721.
Sack, R. O., Carmichael, I.S.E., Rivers, M., and Ghiorso, M. S., 1980, Ferric-ferrous equilibria in natural silicate liquids at 1 bar: Contributions to Mineralogy and Petrology, v. 75, p. 369–376.
Scheidegger, K. F., and Kulm, L. D., 1975, Late Cenozoic volcanism in the Aleutian arc; Information from ash layers in the northeastern Gulf of Alaska: Geological Society of America Bulletin, v. 86, p. 1407–1412.
Scheidegger, K. F., Corliss, J. B., Jazek, P. A., and Ninkovich, D., 1980, Compositions of deep-sea ash layers derived from North Pacific volcanic arcs; Variations in time and space: Journal of Volcanology and Geothermal Research, v. 7, p. 107–137.
Scholl, D. W., Marlow, M. S., MacLeod, N. S., and Buffington, E. C., 1976, Episodic Aleutian Ridge igneous activity; Implications of Miocene and younger submarine volcanism west of Buldir Island: Geological Society of America Bulletin, v. 87, p. 547–554.
Simkin, T., and Smith, J. V., 1970, Minor-element distribution in olivine: Journal of Geology, v. 78, p. 304–325.
Simkin, T., and 5 others, 1981, Volcanoes of the world: Stroudsburg, Pennsylvania, Hutchinson Ross, 233 p.
Simkin, T., Siebert, L., and McClelland, L., 1984, Volcanoes of the World: 1984 Supplement: Smithsonian Institution, Washington, D.C., 33 p.
Simons, F. S., and Mathewson, D. E., 1953, Geology of Great Sitkin Island, Alaska: U.S. Geological Survey Bulletin 1028-B, p. 21–42.
Singer, B. S., and Myers, J. D., 1988, Major and trace element characteristics of lavas from Seguam Island, central Aleutian Islands, Alaska: Geological Society of America Abstracts with Programs, v. 21, p. A56.
—— , 1990, Intra-arc extension and magmatic evolution in the central Aleutian arc, Alaska: Geology, v. 18, p. 1050–1053.
Singer, B. S., Myers, J. D., Linneman, S. R., and Angevine, C., 1989, The thermal history of ascending magma diapirs and the thermal and physical evolution of magmatic conduits: Journal of Volcanology and Geothermal Research, v. 37, p. 273–289.
Singer, B. S., Myers, J. D., and Frost, C., 1992a, Mid-Pleistocene lavas from the Seguam volcanic center, central Aleutian arc; Closed-system fractional crystallization of a basalt to dacite eruptive suite: Contributions to Mineralogy and Petrology, v. 10, p. 87–112.
Singer, B. S., O'Neil, J. R., and Brophy, J. G., 1992b, Oxygen isotope constraints on the petrogenesis of Aleutian Arc magmas: Geology, v. 20, p. 367–370.
Snyder, G. L., 1959, Geology of Little Sitkin Island, Alaska: U.S. Geological Survey Bulletin 1028-H, p. 169–209.
Spence, W., 1977, The Aleutian arc; Tectonic blocks, episodic subduction, strain diffusion and magma generation: Journal of Geophysical Research, v. 82, no. 2, p. 213–230.
Swanson, S. E., Kay, S. M., Brearley, M., and Scarfe, C. M., 1987, Arc and back-arc xenoliths in Kurile-Kamchatka and western Alaska, *in* Nixon, P. H., ed., Mantle xenoliths: New York, John Wiley, p. 303–318.
Tatsumi, Y., Hamilton, D. L., and Nesbitt, R. W., 1986, Chemical characteristics of fluid phase released from a subducted lithosphere and origin of arc magmas; Evidence from high-pressure experiments and natural rocks: Journal of Volcanology and Geothermal Research, v. 29, p. 293–309.
Veniaminov, I., 1840, Notes on the Islands of the Unalashka District: (Black, L. T., and Geoghegan, R. H., translators; Pierce, R. A., ed.): 1984 Limestone Press Edition, Kingston, Ontario, 511 p.
von Drach, V., Marsh, B. D., and Wasserburg, G. J., 1986, Nd and Sr isotopes in the Aleutians; Multicomponent parenthood of island-arc magmas: Contributions to Mineralogy and Petrology, v. 92, p. 13–34.
White, W. M., and Patchett, J., 1984, Hf-Nd-Sr isotopes and incompatible element abundances in island arcs; Implications for magma origins and crust-mantle evolution: Earth and Planetary Science Letters, v. 67, p. 167–185.
Whitney, J. A., and Stormer, J. C., Jr., 1983, Igneous sulfides in the Fish Canyon Tuff and the role of sulfur in calc-alkaline magmas: Geology, v. 11, p. 99–102.
Wilcox, R. E., 1959, Some effects of recent volcanic ash falls with especial reference to Alaska: U.S. Geological Survey Bulletin 1028-N, p. 409–476.
Wilson, F. H., 1985, The Meshik arc; An Eocene to earliest Miocene magmatic arc on the Alaska Peninsula: Alaska Division of Geological and Geophysical Surveys Professional Report 88, 14 p.
Wilson, F. H., Detterman, R. L., and Case, J. E., 1985, The Alaska Peninsula terrane; A definition: U.S. Geological Survey Open-File Report 85-450, 17 p.
Wood, C. A., and Kienle, J., 1990, Volcanoes of North America: Cambridge, Cambridge University Press, 354 p.
Yoder, H. S., 1969, Calcalkalic andesites; Experimental data bearing on the origin of their assumed characteristics, *in* McBirney, A. R., ed., Proceedings of the Andesite Conference: Oregon Department of Geology, Mines and Industry Bulletin, v. 65, p. 77–89.
Yount, M. E., Miller, T. P., Emanuel, R. P., and Wilson, F. H., 1985, Eruption in an ice-filled caldera, Mount Veniaminof, Alaska Peninsula, *in* Bartsch-Winkler, S., and Reed, K. M., eds., The U.S. Geological Survey in Alaska; Accomplishments during 1983: U.S. Geological Survey Circular 945, p. 58–60.

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