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Magmatic and Tectonic Development of the Western Aleutians: An Oceanic Arc in a Strike-Slip Setting

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The Paleogene Basement Series on Attu Island in the Western Aleutians is dominated by tholeiitic basalts chemically similar to mid-ocean ridge basalts (MORB; La/Yb ~ 2, εNd > +10.0, $^{206}Pb/^{204}Pb$ < 18.4). These basalts evolved in "open" magma systems. Contemporaneous rhyolites and albite granites are chemically analogous to silicic volcanic rocks of modern ocean ridges (e.g., Iceland, Galapagos Spreading Center). The occurrence of chemically arc-type tholeiitic basalts, with high Th/La and La/Ta relative to MORB, suggests that a depleted MORB-like mantle source was variably modified by a subduction component. The Attu Basement Series is not allochthonous. Geological and tectonic constraints imply formation in arc-adjacent, transtensional rifts that developed in a strike-slip regime that was established ~43 m.y. ago. Rifting of this kind may have produced much of the Western Aleutian crust between ~43 and 15 Ma. Subsequently, small volumes of crystal-rich andesite and dacite were erupted on Attu Island as well as throughout the Western Aleutians. These strongly calcalkaline rocks (FeO*/MgO~1.1, CaO/Al_2O_3~0.35 at 65% SiO_2) are chemically akin to magnesian andesites of Piip Volcano, a hydrothermally active seamount that overlies small dilational structures within the broadly transpressional regime of the modern western arc. This transpressional setting is inferred to have been established throughout the Western Aleutians approximately 15 m.y. ago. The switch from tholeiitic magmatism in a transtensional regime to strongly calcalkaline magmatism in a transpressional regime may have resulted from clogging of the Aleutian-Kamchatka junction with buoyant, subduction-related terranes. These terranes probably originated to the east and were transported by strike-slip motion along the western arc.

Introduction

An outstanding feature of the modern Aleutian subduction system is the regular decrease in the angle of Pacific-North America convergence that occurs between southern Alaska, where convergence is nearly orthogonal to the trench, and the westernmost arc, where relative Pacific-North America motion is approximately arc-parallel and convergence rates are consequently low (Figure 1). This geometry has produced a well-defined magmatic front and Wadati-Benioff Zone in the Central and Eastern Aleutians, but in the western arc there is no deep or intermediate-depth seismicity, and Pleistocene to Recent magmatism has been volumetrically minor [*Scholl et al.*, 1976; *Borusk and Tsvetkov*, 1982; *Newberry et al.*, 1986; *Boyd and Creager*, 1991; *Tsvetkov*, 1991; *Baranov et al.*, 1991; *Romick et al.*, 1993; *Yogodzinski et al.*, 1993].

The general decrease in orthogonal convergence from east to west along the arc correlates with magma production rates over the past several million years [*Marsh*, 1982]. The relative motions of the Pacific, Eurasian, and North American plates in the North Pacific region have, however, been approximately constant since ~43 Ma [*Engebretson et al.*, 1985]. For this reason, *Vallier et al.* [1993] have suggested that the early magmatic rocks of the Western Aleutian Ridge formed largely prior to 43 Ma when convergence was at a higher angle. This view implies that Western Aleutian arc growth paralleled the earlier stages in the Central and Eastern Aleutians where voluminous magmatism between ~55 and ~37 Ma resulted in rapid growth of the arc crust to nearly its present thickness [*Scholl et al.*, 1970, 1987; *Marlow et al.*, 1973; *Hein and McLean*, 1980].

Contrasting early Tertiary magmatic histories imply, however, that the Western Aleutians has experienced a tectonic evolution unlike that of the central and eastern arc. Early and Middle Tertiary magmatic rocks of the Central and Eastern Aleutians have arc-type geochemical signatures [*Kay and Kay*, 1993], indicating a long-term history of subduction for this part of the arc [*Kay et al.*, 1983; *Rubenstone*, 1984; *McLean*

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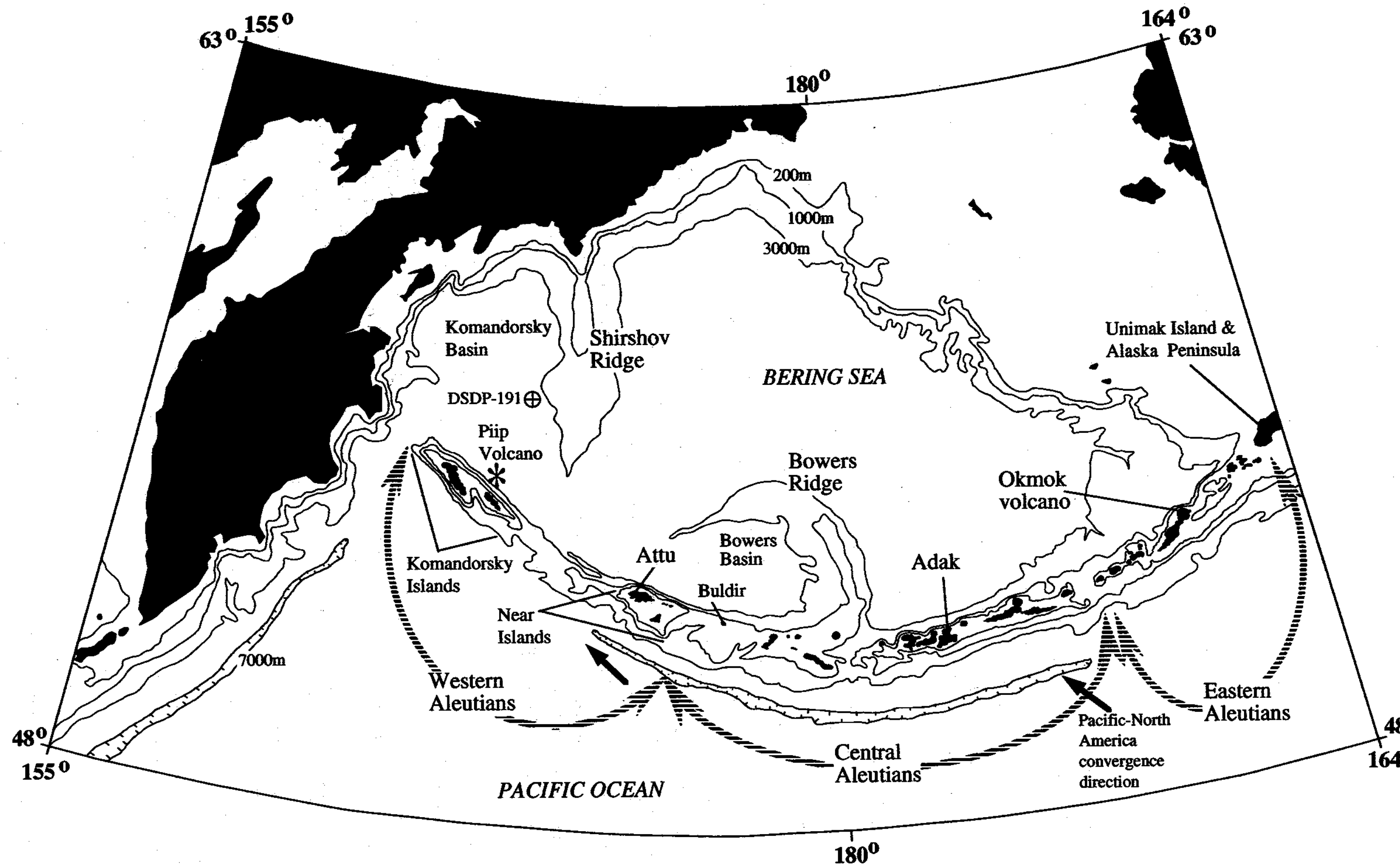

Fig. 1. Insular Aleutian arc and surrounding region. Bathymetry is from *Scholl et al.* [1986]. Modern Pacific-North America convergence in the western and central Aleutians (N49°W, 92 mm/yr and N42°W, 87 mm/yr, respectively) are from *Engebretson et al.* [1985, Table 5]. Piip Volcano in the far western arc is a hydrothermally active seamount [see *Seliverstov et al.*, 1990*a*, *b*; *Baranov et al.*, 1991; *Romick et al.*, 1993; *Yogodzinski et al.*, 1993].

and Hein, 1984; *Hein and McLean*, 1980; *Kay et al.*, 1990; *Kay and Kay*, 1993]. This interpretation is consistent with past plate motions [e.g., *Engebretson et al.*, 1985; *Lonsdale*, 1988] and with the chronology of magmatism in the Alaska-Bering Sea region [e.g., *Scholl et al.*, 1987; *Davis et al.*, 1989]. In contrast, magmatism in the Western Aleutians has undergone dramatic changes from mid-ocean ridge basalt (MORB)-like, to calcalkaline, boninitic, and alkaline compositions [*Borusk and Tsvetkov*, 1982; *Rubenstone*, 1984; *Kay et al.*, 1986; *Shelton*, 1986; *Tsvetkov*, 1991; *Volynets et al.*, 1992; *Romick et al.*, 1993; *Yogodzinski et al.*, 1993]. We infer that the diverse magmatic history of the western arc reflects the diverse tectonic history of the Western Aleutian-Bering Sea region, whereas the geochemical monotony of magmatism in the Central and Eastern Aleutians reflects the relatively simple tectonics of that region.

Exposures of the Western Aleutian Ridge occur on two bathymetrically distinct crustal blocks, the Russian Komandorsky Block in the west and the American Near Islands Block in the east (Figure 1). Here, we use the geochemistry of Eocene through Mio-Pliocene tholeiitic and calcalkaline magmatic successions on Attu Island in the Near Islands as a guide to the changing magmatic sources and evolving tectonics of the Western Aleutian region.

GEOLOGY OF ATTU AND THE NEAR ISLANDS

Attu Island is the largest, most accessible, and best exposed of the Western Aleutian Near Islands. Figure 2 summarizes the geology of Attu after *Gates et al.* [1971, Plate 80]. Five map units composed largely or entirely of volcanic and plutonic rocks were distinguished by *Gates et al.* [1971]. Our results indicate that these units may be combined into two broadly distinctive magmatic series. These are termed the Attu Basement Series and Attu Calcalkaline Series (Figure 2).

The Attu Basement Series

Map units of *Gates et al.* [1971] included within the Attu Basement Series are the earliest Attu rock units (mostly pillow lavas and volcaniclastics), the diabase-gabbro intrusions, and the albite granite-quartz keratophyre intrusions (Figure 2). The age of the Attu Basement Series is poorly constrained, but K-Ar and fossil evidence suggest a formation age between the late Eocene and late Oligocene [*Vallier et al.*, 1993] (also see discussion below).

The oldest and volumetrically most important units in the Attu Basement Series are pillow lavas, volcanic breccias, and volcaniclastic rocks, with interbedded marine sedimentary

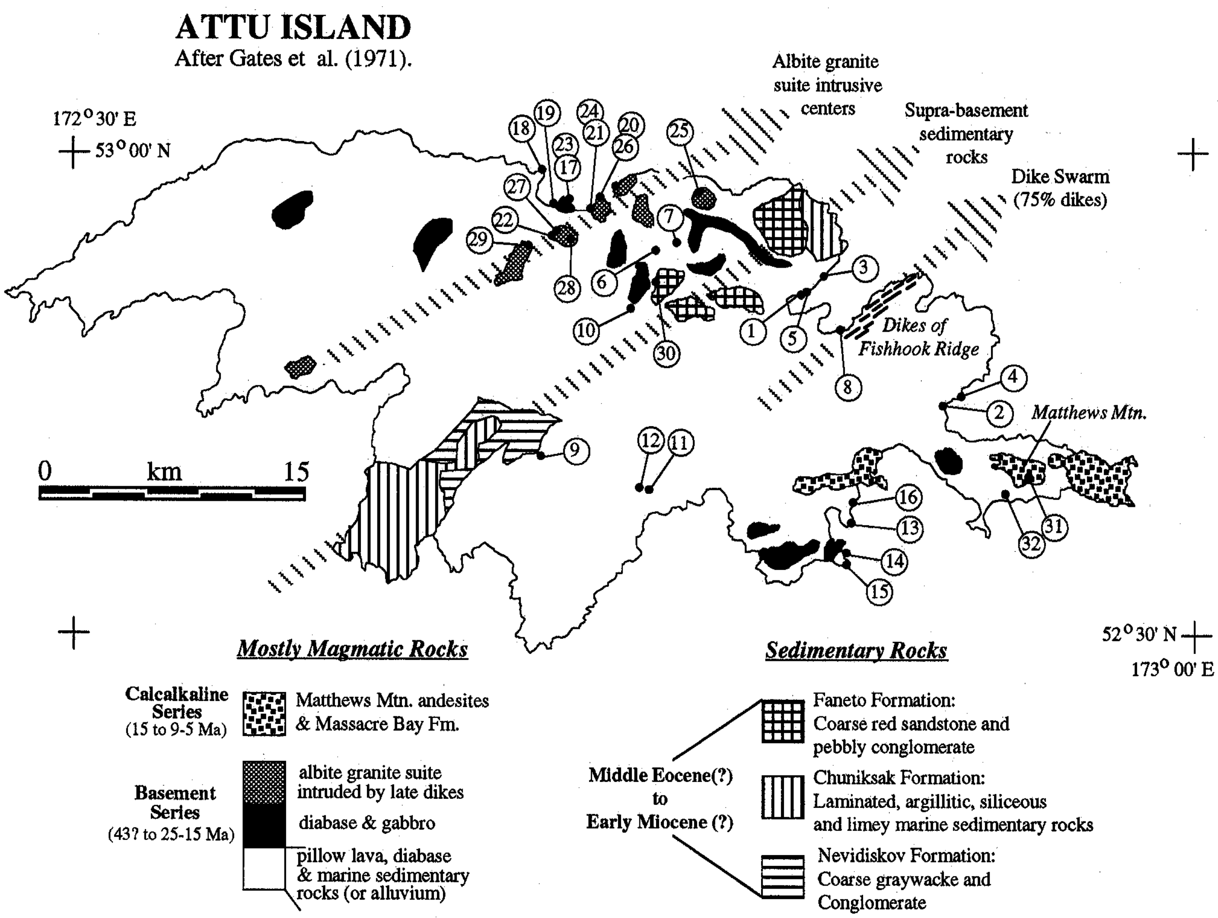

Fig. 2. Geologic sketch map of Attu Island. The Basement Series consists of the oldest map units from *Gates et al.* [1971]: pillow lavas, diabase-gabbro, albite granite, and quartz keratophyre. The Calcalkaline Series includes the Matthews Mountain andesites and volcaniclastics of the Massacre Bay Formation. Intrusive rocks of the Attu Basement Series that crosscut the Faneto Formation imply an older age for the Faneto Formation than assigned by *Vallier et al.* [1993]. See text for discussion of other age assignments. Large, dashed arrows highlight northeast trending belts thought to reflect the structural and magmatic alignment during formation of the Attu Basement Series. Sample locations are keyed to Tables 1*a* and 1*b*; see also Table 2.

TABLE 1*b*. Attu Whole Rock Major and Trace Element Analyses

	Map Site / Specimen															
	Late Dikes						Albite Granite-Suite								Calcalkaline	
	17-D/ AT11	18-D/ AT50	19-D/ AT16	20-D/ AT27	21-D/ AT22	22-D/ AT21	23-G/ AT6	24-IC/ AT23	25-R/ AT65	26-R/ AT29	27-R/ AT19	28-AG/ AT3	29-AG/ AT56	30-R/ AT143	31-A*/ AT8032	32-A/ AT83
SiO_2	49.90	50.63	48.70	49.75	53.17	58.35	47.48	60.92	72.30	72.99	73.79	74.51	76.72	76.99	58.20	56.94
TiO_2	2.16	2.29	2.95	3.00	1.85	2.04	4.44	1.34	0.30	0.43	0.29	0.46	0.22	0.25	0.65	0.61
Al_2O_3	15.50	15.48	15.58	14.64	16.60	16.24	15.35	17.21	13.80	14.45	13.79	12.93	12.65	12.65	18.66	18.03
FeO*	10.42	10.68	11.78	13.02	9.75	8.51	13.30	6.17	2.37	1.84	2.71	2.12	1.16	1.41	5.72	6.13
MnO	0.17	0.18	0.21	0.22	0.17	0.12	0.25	0.10	0.04	0.03	0.04	0.03	0.02	0.02	0.15	0.12
MgO	6.57	6.41	7.50	4.38	5.31	3.67	4.91	1.77	0.14	0.39	0.24	0.30	0.22	0.16	4.44	4.59
CaO	8.60	8.60	7.33	7.72	5.87	3.94	10.21	3.87	0.45	0.79	0.39	0.38	0.71	0.49	6.83	6.67
Na_2O	3.84	3.95	4.32	4.93	5.80	6.12	3.35	7.53	5.89	7.15	6.84	6.43	4.80	5.54	4.03	3.47
K_2O	1.57	1.18	0.53	0.71	1.01	1.87	0.54	0.83	3.31	1.59	1.61	2.64	3.14	2.19	1.12	1.56
Total	98.73	99.40	98.90	98.37	99.53	100.86	99.83	99.74	98.60	99.66	99.70	99.80	99.64	99.70	99.80	98.12
La	8.66	9.30	11.90	11.4	13.7	23.2	5.11	16.4	32.7	26.8	28.0	36.5	20.4	24.6	8.80	6.85
Ce	24.5	25.1	33.9	32.0	34.6	50.9	14.0	41.0	79.8	69.1	59.5	93.2	53.0	54.0	20.1	15.8
Nd	.	18.2	24.7	22.3	23.2	.	.	32.6	39.1	41.8	.	58.9	24.8	26.4	12.7	10.8
Sm	5.33	5.52	7.69	6.83	6.11	9.41	4.03	9.42	9.63	9.87	10.4	15.4	5.84	5.34	2.72	2.52
Eu	1.87	1.78	2.37	2.24	1.98	2.86	1.82	5.56	1.54	2.36	2.31	3.23	0.80	0.69	0.78	0.74
Tb	1.11	1.22	1.67	1.37	1.20	1.80	0.87	1.67	1.85	1.80	1.93	2.84	1.22	0.87	0.40	0.37
Yb	3.46	3.93	5.32	4.35	3.83	6.64	2.04	3.60	8.65	8.20	9.15	12.60	6.37	5.05	1.56	1.50
Lu	0.480	0.550	0.750	0.600	0.540	0.990	0.280	0.490	1.28	1.19	1.42	1.92	0.900	0.700	0.210	0.215
Sr	531	399	229	158	.	.	.	468	.	.	.	.	.	.	.	.
Ba	597	246	82	97	212	308	91	317	349	276	236	279	304	335	375	412
Cs	0.15	0.23	0.13	0.02	0.36	0.03	0.15	0.05	0.17	0.02	0.06	0.15	0.15	0.02	0.10	0.17
U	0.09	0.18	0.23	0.23	0.49	0.91	.	0.11	1.83	1.28	1.36	2.37	1.88	2.13	0.71	0.53
Th	0.43	0.45	0.74	0.57	1.17	1.71	0.18	0.57	4.25	2.47	3.10	5.19	5.02	5.23	1.14	0.83
Hf	3.79	4.17	6.28	4.61	5.17	7.86	1.82	2.10	14.19	11.80	13.70	16.20	5.94	5.18	2.69	2.43
Ta	0.54	0.52	0.87	0.61	0.65	0.99	0.56	0.87	2.81	1.82	2.33	3.01	2.43	1.61	0.20	0.12
Sc	45	41	37	38	30	26	49	20	7	6	12	9	2	2	19	19
Cr	137	124	119	8	38	34	5	3	2	5	13	8	5	13	105	111
Ni	34	46	69	14	37	12	1	12	1	4	6	7	2	2	40	51
Co	40	40	43	38	36	24	34	22	20	21	10	9	4	6	21	22

Map is keyed to Figure 2; letter refers to rock type or occurrence (D is dike, G is gabbro/diorite, IC is intermediate composition plutonic rock, AG is albite granite, R is rhyolite, A is hornblende andesite). Major element analyses are by electron microprobe; trace element analyses are by instrumental neutron activation analysis. Technique descriptions and standard analyses are given by *Kay and Kay* [1988] and *Kay et al.* [1987]. Analytical precision on trace elements is given by *Romick et al.* [1992].

*Analysis from *Rubenstone* [1984].

TABLE 1*a*. Attu Island Whole Rock Major and Trace Element Analyses

	Map site/Specimen															
	MORB-like												Arc-like			
	1-P/ HO925B	2-P*/ SB1A	3-P*/ HO914B	4-P*/ SB5A	5-P*/ HO922	6-P/ AT61	7-P/ AT73	8-P/ HO962A	9-M/ AT114	10-M/ AT144	11-M/ AT101	12-P/ AT100	13-P*/ CP98	14-P*/ MP2	15-P*/ MP10	16-P/ CP916C
SiO_2	49.37	49.70	49.04	51.40	49.79	48.50	50.58	49.47	53.28	51.52	52.77	55.55	53.59	53.46	52.85	56.53
TiO_2	0.96	1.46	1.53	1.64	1.83	1.74	1.98	1.64	1.08	1.91	1.40	1.23	0.87	0.96	0.99	0.90
Al_2O_3	18.67	15.49	18.45	15.31	15.47	16.25	16.75	18.00	15.72	16.35	15.61	15.52	16.06	16.17	16.79	16.65
FeO*	8.84	8.94	8.60	9.03	10.03	9.29	9.47	7.48	11.61	11.70	12.45	11.15	11.33	12.01	12.86	10.15
MnO	0.13	0.12	0.18	0.21	0.16	0.16	0.15	0.14	0.17	0.19	0.20	0.16	0.22	0.18	0.26	0.17
MgO	9.96	8.52	8.24	8.79	8.79	7.02	7.11	5.48	5.12	4.96	4.53	3.94	5.44	5.70	4.79	4.33
CaO	6.81	11.47	10.30	8.17	8.04	11.27	9.42	13.34	9.00	7.94	8.38	5.12	5.76	4.73	6.61	7.03
Na_2O	3.78	3.82	3.38	3.37	3.84	3.60	4.66	3.08	2.90	4.84	3.45	6.59	4.70	4.70	5.95	3.27
K_2O	0.80	0.23	0.20	0.76	1.27	0.46	0.28	0.79	1.12	0.32	0.38	.028	1.12	2.83	0.11	0.25
Total	99.32	99.75	99.92	98.68	99.22	98.29	100.40	99.42	100.00	99.73	99.17	99.54	99.16	100.88	101.38	99.28
La	2.33	3.90	6.26	6.91	6.10	5.22	7.36	9.64	2.40	3.98	3.09	3.07	4.51	4.89	6.53	6.36
Ce	6.73	12.7	18.4	20.3	18.8	15.8	19.6	22.1	6.36	11.2	8.75	8.54	11.3	12.9	16.3	15.3
Nd	4.5	10.1	13.4	14.5	14.9	12.4	15.6	15.0	6.3	9.5	.	8.1	.	.	.	.
Sm	2.08	3.22	3.75	4.37	4.51	4.16	5.10	4.25	2.22	3.43	3.04	3.02	2.75	2.99	3.26	3.47
Eu	0.80	1.19	1.36	1.54	1.57	1.41	1.78	1.30	0.75	1.41	1.09	0.99	0.93	0.95	1.27	1.09
Tb	0.50	0.81	0.94	1.09	1.04	0.95	1.11	0.81	0.58	0.87	0.80	0.79	0.57	0.63	0.64	0.79
Yb	1.69	2.47	2.73	3.30	3.23	3.20	4.28	2.75	2.37	2.98	2.87	3.03	2.00	2.26	2.22	2.69
Lu	0.253	0.385	0.422	0.518	0.487	0.430	0.680	0.410	0.358	0.420	0.422	0.439	0.297	0.332	0.308	0.359
Sr	313	183	229	206	267	414	355	.	172	286	231	111	255	494	235	269
Ba	104	40	35	65	204	125	38	87	167	55	84	31	392	1526	16	42
Cs	0.18	0.19	0.28	0.36	0.30	0.86	0.14	0.14	0.13	0.18	0.12	0.05	0.06	0.30	.	0.07
U	.	0.12	0.31	0.07	0.77	0.09	0.30	0.43	0.10	.	0.05	0.06	0.15	0.11	0.23	0.06
Th	0.10	0.22	0.38	0.44	0.30	0.23	0.16	0.90	0.24	0.14	0.19	0.16	0.11	0.49	0.44	0.18
Hf	1.41	2.47	3.00	3.54	3.41	3.21	4.00	3.34	1.60	2.22	2.06	2.16	1.58	1.83	1.61	1.62
Ta	0.12	0.18	0.37	0.30	0.26	0.31	0.46	1.20	0.11	0.30	0.27	0.13	0.05	0.07	0.06	0.04
Sc	35	36	36	3	37	38	42	32	42	36	36	29	40	43	43	39
Cr	269	408	363	407	342	348	263	328	26	18	13	13	18	15	32	24
Ni	103	185	187	153	168	171	127	206	14	7	13	10	9	14	17	16
Co	43	44	41	38	44	45	44	40	37	28	40	29	36	38	43	32

Map is keyed to Figure 2; letter refers to rock-type or occurrence (P is pillow basalt, M is massive diabase). Major element analyses are by electron microprobe, trace element analyses are by instrumental neutron activation analysis. Technique descriptions and standard analyses are given by *Kay and Kay* [1988] and *Kay et al.* [1987]. Analytical precision on trace elements is given by *Romick et al.* [1992].

*Analysis from *Rubenstone* [1984].

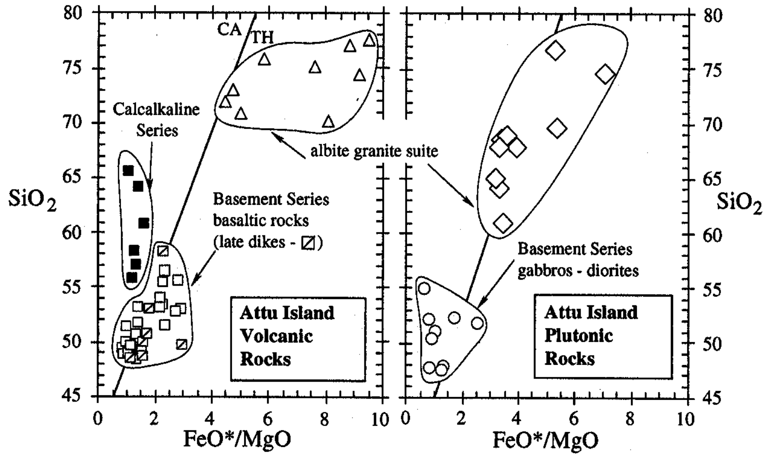

Fig. 3. SiO_2 versus FeO*/MgO for magmatic rocks of Attu Island. Note that the Basement Series rocks are bimodal with a compositional gap between 60 and 70% SiO_2. Basement Series gabbros and diorites include samples from the dominantly silicic intrusive centers of the albite granite suite (e.g., sample AT6, Table 1*b*; also see text). Data from Aggatu Island (large, unlabeled Near Island in Figure 1) are included in the Attu Calcalkaline Series. Calcalkaline - tholeiitic line from *Miyashiro* [1974]. Data from Table 1 here, *Gates et al.* [1971], *Shelton* [1986], and unpublished Cornell data.

pattern of the Komandorsky Basin basalt is like that of the Attu MORB-like tholeiites suggesting that this is a regional mantle characteristic (Figure 4).

The arc-like tholeiites, the second type of Attu basalt, have all of the trace element characteristics expected in subduction-related volcanic rocks. They have relatively low concentrations of HREE (e.g., low Yb relative to Cr; see Figure 7), and they are flat to slightly enriched in the LREE with high relative Th compared to the MORB-like samples (Figures 4 and 5). Most characteristically, the Attu arc-like tholeiites have low relative concentrations of the HFSE (Figures 4*b*, 5*c*, and 6).

Isotopic characteristics of the Attu MORB-like and arc-like tholeiites are broadly consistent with their trace element and petrological affinities. All of the Attu tholeiites have εNd within the range of North Pacific MORB (Figure 8), but on average, the MORB-like samples are more radiogenic (+10 to +12) than the arc-like samples (+9 to +10). High $^{87}Sr/^{86}Sr$ in the Attu rocks (> 0.7030; see Figure 8) relative to NMORB are thought to reflect postmagmatic alteration [*Rubenstone et al.*, 1982]. In contrast, Pb isotope ratios cluster in the nonradiogenic region of the North Pacific MORB field (Figure 9) and are consistent with geochemical parameters that are resistant to hydrothermal exchange (e.g., high $^{207}Pb/^{204}Pb$ and Th/Ta with low $^{143}Nd/^{144}Nd$ in the arc-like rocks). The linear trend of the Attu basalts on Pb-Pb diagrams (e.g., Figure 10) suggests that disruption of $^{238}U/^{204}Pb$ by alteration has been relatively minor because 30-40 m.y. of radioactive decay have not produced highly variable $^{206}Pb/^{204}Pb$ [e.g., *Hamelin et al.*, 1984]. The Pb isotope ratios of the MORB-like and arc-like tholeiites thus appear to be magmatic values.

Geochemistry of the Albite Granite Suite

Albite granites (SiO_2=67-77%) and compositionally equivalent rhyolites have high Na_2O (4.3-7.2%), low CaO (0.2-1.3%), and variable K_2O (1.8-3.3%) which generally increases with increasing SiO_2. These rocks have high incompatible trace element concentrations (La up to ~100X chondrites), they are slightly Th and LREE-enriched (chondrite-normalized La/Sm ~2.0-2.5), and have flat HREE patterns (normalized Tb/Yb ~1.0, Figure 11). Large negative Eu anomalies in the albite granites and rhyolites increase with increasing SiO_2 and REE abundance (up to SiO_2~75%, Eu/Eu*=0.60, and La=97X chondrites). At higher SiO_2 (up to 77%), Eu anomalies continue to increase (to Eu/Eu*=0.40), whereas REE and Hf concentrations drop sharply and the most incompatible elements (Th, U, etc.) continue to increase or remain unchanged (Figure 11 and Table 1*b*).

The rhyolites and albite granites have Nd and Sr isotopic compositions that are broadly similar to the MORB-like tholeiites. They have εNd (8.6-9.6) that is on average slightly less radiogenic than North Pacific MORB, and $^{87}Sr/^{86}Sr$ that are high relative to εNd (Figure 8). As in the basaltic rocks, high $^{87}Sr/^{86}Sr$ in the rhyolites and albite granites was probably produced by exchange with seawater Sr. The Pb isotopic compositions of the silicic rocks are distinctly more radiogenic than the MORB-like basalts; $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ in the rhyolites and albite granites overlap the most radiogenic of the MORB-like and arc-like tholeiites, while $^{206}Pb/^{204}Pb$ in the silicic rocks extends to much higher values (Figure 10 and Table 3).

Intermediate composition rocks (SiO_2 58-66%) have extremely high Na_2O (6.3-8.1%), relatively low CaO (1.3-1.4%), and highly variable K_2O (0.16-2.0%). Most of the intermediate composition plutonic rocks have either no Eu anomaly, a small positive Eu anomaly (Eu/Eu*= 1.1-1.2), or in one case a large positive Eu anomaly (Eu/Eu*=1.8; Figure 12). Isotopic analyses are not available for the intermediate composition plutonic rocks. On the basis of their mineralogy, texture, and geochemistry, the intermediate composition plutonic rocks appear to be crystal cumulates.

Geochemistry of the Late Dikes

The late dikes are also tholeiitic (high FeO*, FeO*/MgO), and they fall along approximately the same fractionation trends as most of the Basement Series MORB-like tholeiites (e.g., Figure 7). The late dikes are however, more evolved, ranging to ferrobasalt (TiO_2 up to 3.0%, FeO* up to 13%, FeO*/MgO up to 2.8), and andesitic compositions (sample AT21, Table 1) with higher incompatible element abundances

TABLE 2. Sample Locations and Petrography

Sample	Structure	Rock Type	Phenocrysts	Latitude	Longitude	Elevation m (feet)
HO925B	pillow lava	basalt	pl	52°56.14'N	173°07.88'E	0
SB1A	pillow lava	basalt	ol	52°52.96'N	173°16.10'E	0
HO914B	pillow lava	basalt	aphyric	52°56.14'N	173°07.88'E	0
SB5A	pillow lava	basalt	ol	52°52.91'N	173°16.21'E	0
H922	pillow lava	basalt	pl-ol-cpx	52°56.22'N	173°08.40'E	0
AT61	pillow lava	basalt	aphyric with cpx megacrysts	52°57.95'N	173°00.10'E	671 (2200)
AT73	pillow lava	basalt	pl	52°58.32'N	173°01.75'E	701 (2300)
H962A	pillow lava,	basalt	aphyric	52° 55.04'N	173°10.05'E	0
AT114	massive diabase	basaltic andesite	pl-cpx	52°51.20'N	172°55.30'E	640 (2100)
AT144	massive diabase	basalt	aphyric	52°55.45'N	172°59.75'E	335 (1100)
AT101	massive diabase	basalt	aphyric	52°50.10'N	173°00.45'E	533 (1750)
AT100	massive diabase	basaltic andesite	aphyric	52°50.10'N	173°00.35'E	457 (1500)
CP98	pillow lava	basalt	pl-cpx-olv	52°49.08'N	173°10.83'E	0
MP2	pillow lava	basalt	pl-ol	52°48.40'N	173°10.44'E	0
MP10	pillow lava	basalt	pl-ol	52°47.82'N	173°08.67'E	0
CP916C	pillow lava	basalt	pl-cpx	52°49.59'N	173°10.91'E	0
AT11	dike	basalt	pl-cpx	52°58.60'N	172°56.00'E	0
AT50	dike	basalt	pl-cpx-olv	52°59.65'N	172°54.60'E	0
AT16	dike	basalt	pl-cpx	52°58.62'N	172°55.40'E	0
AT27	dike	basalt	aphyric	52°58.80'N	172°57.60'E	0
AT22	dike	basaltic andesite	pl-cpx-olv	52°58.55'N	172°57.45'E	0
AT21	dike	andesite	aphyric with trace amph-ox.	52°57.60'N	172°54.70'E	472 (1550)
AT6	stock	gabbro	pl-cpx-ox-amph	52°58.52'N	172°56.35'E	0
AT23	stock	plutonic rock	pl-ox-amph-qtz	52°58.55'N	172°57.45'E	0
AT65	dike	rhyolite	pl-ox	52°59.40'N	173°01.10'E	76 (250)
AT29	lava	rhyolite	pl-amph-ox	52°58.80'N	172°57.60'E	0
AT19	lava	rhyolite	pl-ox	52°57.85'N	172°55.00'E	457 (1500)
AT3	stock	albite granite	pl-qtz-cpx-ox	52°57.60'N	172°56.25'E	472 (1550)
AT56	stock	albite granite	pl-qtz-ox-biot-amph	52°56.65'N	172°53.00'E	76 (250)
AT143	sill intruding Faneto Formation	rhyolite	pl-qtz.	52°56.15'N	173°01.30'E	640 (2100)
AT8032	stock	andesite	pl-amph-rich	52°51.04'N	173°18.5'E	vicinity of Alexai Pass
AT83	dike	andesite	pl-amph-rich	52°49.80'N	172°17.50'E	0

Mineral abbreviateions are cpx (clinopyroxene), olv (olivine), pl (plagioclase), ox (Fe-Ti oxides), and amph (amphibole). Phenocryst phases in volcanic rocks and minerals in plutonic rocks are listed in order of decreasing abundance. The former presence of olivine is inferred from alteration products.

(e.g., La = 20-40X chondrites) and lower compatible element abundances. The late dikes have, on average, higher La/Sm, $^{207}Pb/^{204}Pb$, and lower average εNd than the older MORB-like tholeiites (Figures 4-10). None of the late dikes have the HFSE-depleted trace element signature of of the arc-like basement tholeiites.

Geochemistry of the Matthews Mountain Andesites and Other Western Aleutian Calcalkaline Rocks

The Calcalkaline Series rocks are distinguished from those of the Basement Series by their silicic-to-intermediate composition (SiO_2=55-65%) and low FeO*/MgO (1.0-1.5,

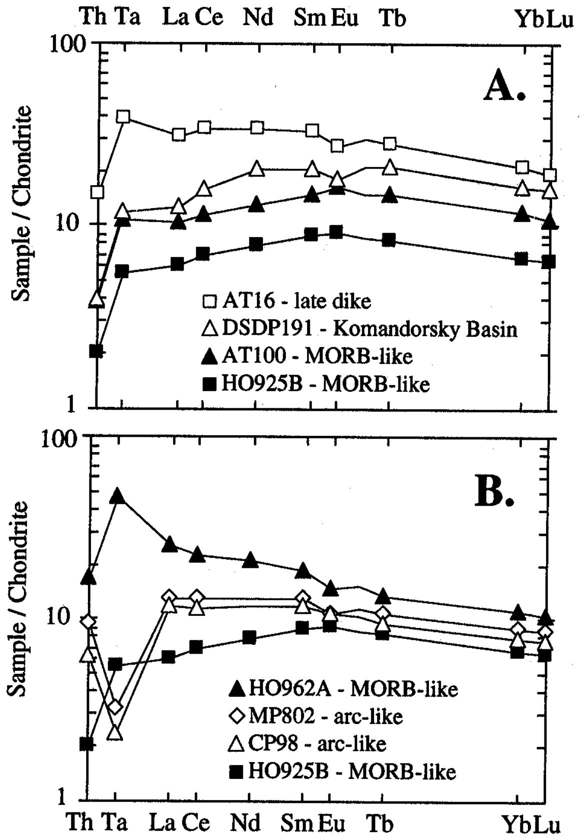

Fig. 4. Chondrite-normalized rare earth elements (REE), Th, and Ta in mafic volcanic rocks of the Attu Basement Series and in Komandorsky Basin basalt (DSDP-191; analysis from *Rubenstone* [1984]). In Figure 4*a*, note that mid-ocean ridge basalt (MORB)-like samples are light REE-depleted, similar to the Komandorsky Basin basalt, whereas the late dike sample has slightly higher La/Sm. Figure 4*b* includes a light REE-enriched MORB-like sample similar to enriched MORB (see also Figure 5). Note also that arc-like samples have high Th and low Ta but cannot be distinguished by their REE patterns alone. Normalizing values are Leedy Chondrite: Th (0.050), Ta (0.022), La (0.378), Ce (0.976), Nd (0.716), Sm (0.230), Eu (0.0866), Tb (0.0589), Yb (0.249), and Lu (0.0387).

Figure 3). As noted above, the Near Islands Calcalkaline Series is petrographically diverse. The discussion here focuses principally on the Matthews Mountain andesites which are located on Attu (Tables 3 and 4). These rocks appear to be broadly representative of middle Miocene magmatic rocks throughout the Near Islands and on the Komandorsky Islands of the westernmost arc (also see discussion below).

The Matthews Mountain andesites have geochemical characteristics similar to calcalkaline rocks of the modern Central and Eastern Aleutian arc. They are relatively rich in large ion lithophile and LREE (La/Yb~5, Ba/La~40-60) and have low relative HFSE abundances (Figures 6 and 13). The Matthews Mountain andesites are, however, distinct from the Central and Eastern Aleutian suite in two important ways. First, incompatible element abundances (especially Th and HREE) are frequently lower than in calcalkaline samples of equivalent SiO_2 content from the central and eastern arc (Figure 14); second, the Matthews Mountain andesites are isotopically more similar to MORB. The MORB-like isotopic character is seen most clearly in Pb-Pb plots where the Matthews

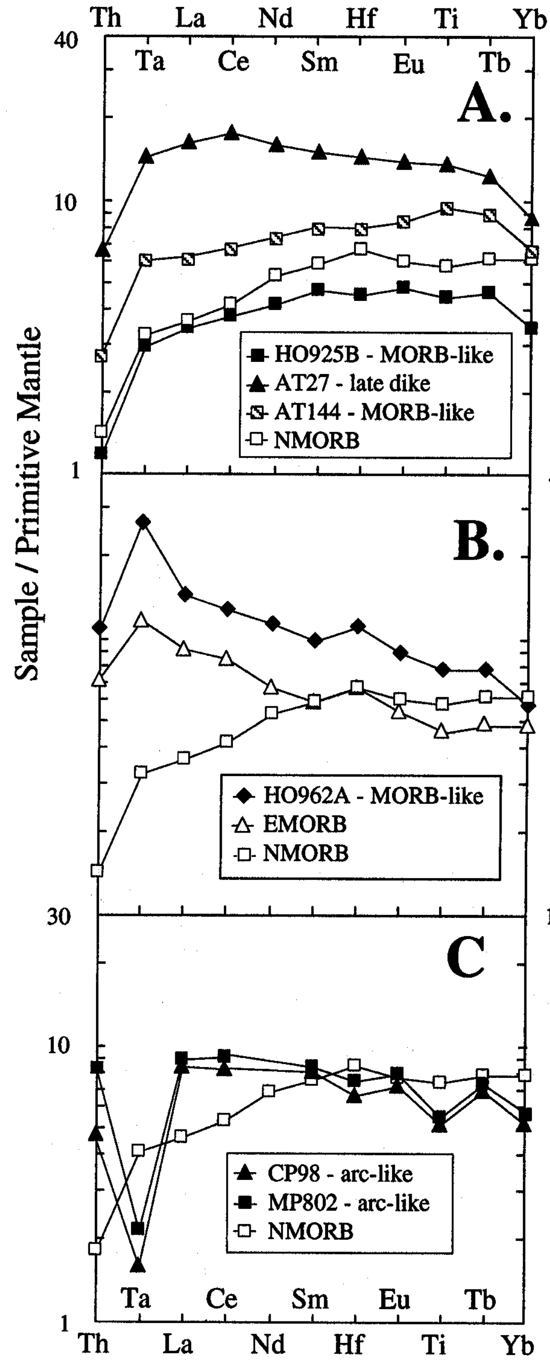

Fig. 5. Immobile-incompatible elements in Attu Basement Series basalts normalized to primitive mantle. In Figure 5*a*, note the broad similarity of the late dikes to the mid-ocean ridge basalt (MORB)-like samples and to normal MORB. Note also that sample AT144, which has some characteristics of the arc-like samples (Cr~20, Yb~3 ppm; see Figure 7 and Table 1*a*) is MORB-like with respect to its incompatible trace element characteristics. In Figure 5*b*, note the similarity of pillow lava HO962A and enriched MORB. In Figure 5*c*, note the low relative concentrations of Ti, Hf, and Ta and the high relative concentrations Th in the arc-like Attu basalts. MORB and normalizing values are from *Sun and McDonough* [1989]. Eu value is interpolated between Sm and Tb from REE patterns for Attu samples with small negative Eu anomalies.

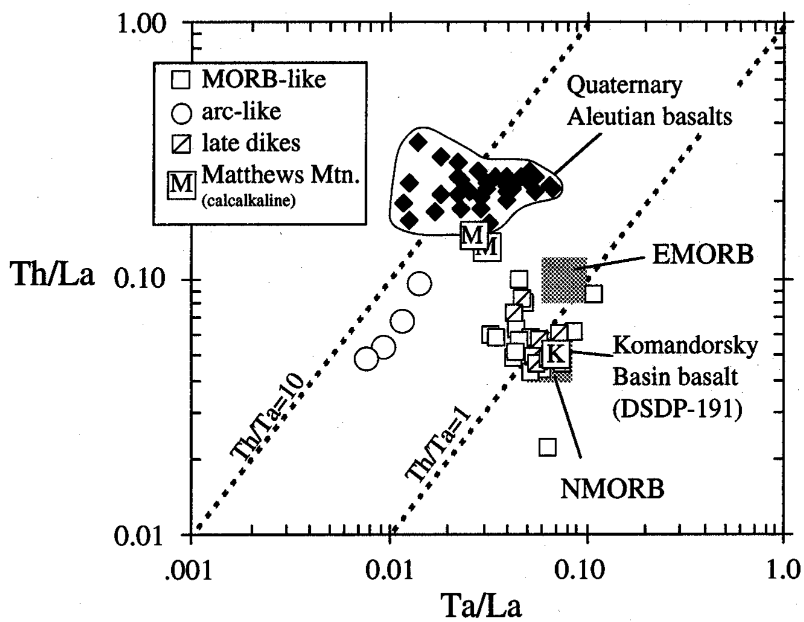

Fig. 6. Th/La versus Ta/La for Attu rocks compared to mid-ocean ridge basalts (MORB) and Quaternary basalts of the Central and Eastern Aleutians. Note similarity of Attu MORB-like tholeiites and late dikes to the Komandorsky Basin basalt and to MORB. Note also similarity of Matthews Mountain andesites to Aleutian basalts. Attu arc-like basalts have Ta/La ratios like Quaternary Aleutian suite but lower relative Th concentrations. MORB values from *Sun and McDonough* [1989]. Data are from Table 1, and *Kay and Kay* [1993].

Mountain andesites are indistinguishable from North Pacific MORB (Figure 10). Similarly, εNd in the Matthews Mountain andesites overlap with the most radiogenic of the modern Aleutian suite and with the least radiogenic of North Pacific MORB (Figure 8).

MORB-like isotopic characteristics and low relative Th contents are not typical of calcalkaline magmatic rocks of the modern Central and Eastern Aleutian arc. These characteristics are, however, seen in Miocene volcanic and subvolcanic rocks of the far Western Aleutian Komandorsky Islands [*Borusk and Tsvetkov*, 1982; *Tsvetkov*, 1991] and in Recent volcanic rocks of Piip Volcano [*Yogodzinski et al.*, 1993; *Romick et al.*, 1993] (locations in Figure 1). Post middle Miocene Western Aleutian magmatic rocks (Near Islands, Komandorsky Islands, and Piip Volcano) have lower FeO*/MgO and CaO/Al_2O_3 than most calcalkaline rocks of the modern Central and Eastern Aleutian arc, and in this respect the Western Aleutian rocks form a more strongly calcalkaline series (Figure 15).

DISCUSSION AND GEOCHEMICAL MODELING

Evidence for a Rift Tectonic Setting

Small geochemical differences notwithstanding, source characteristics of the Attu Basement Series rocks are unmistakably those of NMORB. It is thus an important first-order conclusion that during middle Eocene through late Oligocene time the mantle beneath the Western Aleutians was a depleted peridotite similar to the source of modern MORB.

Beyond similarities in source chemistry, the Attu Basement Series rocks appear to have formed by processes physically similar those that operate at modern oceanic or back-arc spreading centers. The magmatic sequence on Attu began with the eruption of voluminous, chemically MORB-like tholeiitic basalts, progressed to the formation of highly evolved silicic differentiates (albite granites), and ended with the intrusion ferrobasaltic and andesitic dikes (late dikes). This sequence is similar to that observed at propagating oceanic rifts [*Christie and Sinton*, 1981] and is reasonable within the context of physical-chemical models of ocean ridge magma chambers [*Sinton and Detrick*, 1992].

A rift tectonic setting is also implied by the style of metamorphism on Attu. Most of the Basement Series rocks are subgreenschist in grade, but *Rubenstone* [1984] and *Rubenstone et al.* [1982] noted that metamorphic gradients are were locally high, reaching amphibolite facies at the Fishook Ridge dike swarm. This metamorphic style contrasts with that in basement assemblages of the Central and Eastern Aleutians (e.g., Adak), where metamorphism resulted from fracture-controlled hydrothermal circulation of meteoric water (low $\delta^{18}O$) driven by heat from the younger plutonic bodies [*Perfit and Lawrence*, 1979; *Rubenstone et al.*, 1982; *Kay*, 1983]. The style of metamorphism on Adak is like that expected in an emergent island arc. The style on Attu is similar to that produced by dynamic hydrothermal metamorphism beneath oceanic rift valleys [*Bonatti et al.*, 1975], and is common in ophiolites. Indeed, the abundance of pillow basalts and laminated marine sedimentary rock in the Near Islands prompted *Wilcox* [1956] to comment that except for the absence of serpentinized peridotite, the Attu Basement Series was similar to an ophiolite. Such characteristically oceanic assemblages are not found in early and middle Tertiary sequences of the Central and Eastern Aleutians.

If the Attu crust formed by a processes broadly analogous to seafloor spreading or back-arc rifting, then crystal-liquid fractionation in the Attu tholeiites may also follow the characteristic evolutionary behavior of modern ocean-ridge systems. A plot of incompatible element enrichment (relative to the most primitive sample) versus FeO*/MgO shows that

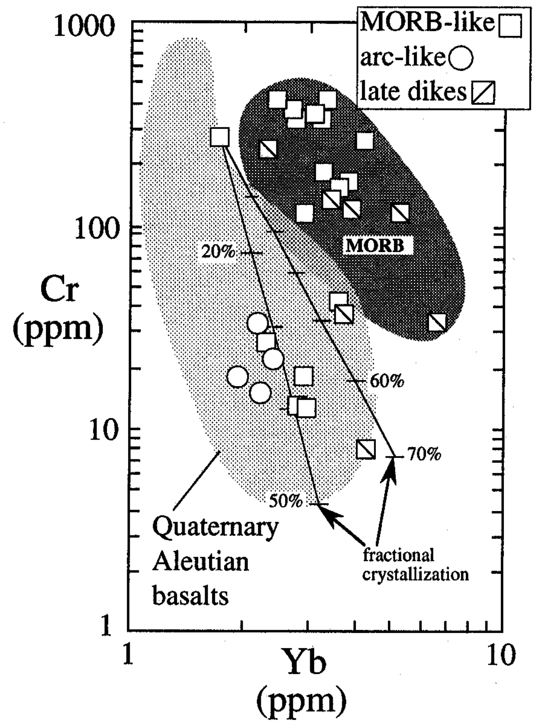

Fig. 7. Cr versus Yb in Attu basalts. Low-Cr mid-ocean ridge basalt (MORB)-like samples (<30 ppm) have isotopic and incompatible trace element characteristics like MORB (e.g., sample AT144 in Figure 5*a*). The arc-like basalts have the lowest Yb relative to Cr among the Attu basalts. Closed-system Rayleigh fractionation lines calculated for bulk solid/liquid distribution coefficients (D) DYb=0.1, DCr=7 (lower curve), and DCr=4 (upper curve). Aleutian basalt field from *Kay and Kay* [1993] and references therein. MORB field is approximated from published and unpublished analyses from the Mid-Atlantic Ridge, East Pacific Rise, and Juan de Fuca and Gorda ridges.

most of the MORB-like tholeiites and late dikes do not follow closed system, fractional crystallization paths (Figure 16). In general, the Attu trend is one of rapidly increasing incompatible element concentration relative to FeO*/MgO. The compatible and incompatible elements in the Attu MORB-like tholeiites and late dikes are thus decoupled. This kind of behavior is common in MORB [e.g., *Walker et al.*, 1979], and is frequently attributed to evolution in an open or periodically recharged magma chamber [*O'Hara*, 1977]. Similar trends may be produced by in situ crystallization, where magmas evolve by liquid exchange between the hot, convecting part of the chamber and its cooler, partly crystalline margins [e.g., *Langmuir* 1989]. The actual fractionation mechanism is less important than the observation that incompatible elements are strongly enriched relative to FeO*/MgO in the Attu rocks. This pattern is not typical of tholeiitic volcanos of the modern Aleutian arc (see Okmok caldera sequence in Figure 16). Only the low-Cr MORB-like Attu basalts (Cr<40 ppm; see Figure 7) have the relatively low enrichment factors expected from simple closed system evolution (Figure 16) .

Origin of the Albite Granites

Within the context of an oceanic or back-arc rift analogue for the Attu Basement Series, the origin of the albite granite suite remains unexplained. Examples of silicic rocks in oceanic rift settings include the rhyolites and rhyodacites from the Galapagos Spreading Center [*Clague et al.*, 1981], quartz diorites and trondhjemites from the Mid-Atlantic Ridge [*Aldiss*, 1981], and Icelandic rhyolites [*O'Nions and Gronvold*, 1973; *Wood*, 1978; *Shimokawa and Masuda*, 1972]. Moreover, bimodal associations are a common feature of back-arc spreading centers in the Western Pacific [see *Lonsdale and Hawkins*, 1985; *Ishizuka et al.*, 1990; *Hochstaedter et al.*, 1990; *Fryer et al.*, 1990]. Silicic end-members in all of these settings are similar to the Attu rhyolites in that they have high incompatible element abundances, large negative Eu anomalies, and relatively unfractionated REE patterns. Similar REE characteristics are also seen in subduction-related tholeiitic systems (e.g., dacitic rocks from Okmok Volcano, Eastern Aleutians), but the isotopic characteristics of the Attu rocks are more akin to MORB than to normal arc volcanic rocks.

The origin of rhyolites in oceanic bimodal assemblages remains a matter of debate [e.g., *Sigmarsson et al.*, 1991; *MacDonald et al.*, 1990], but many authors favor extreme fractional crystallization of basaltic parents. In this regard, the Attu late dikes are of interest, because they include Fe-Ti rich compositions, similar to those associated with silicic volcanism at oceanic rifts (e.g., Iceland and Galapagos Spreading Center). The late dikes range from normal tholeiitic basalts (FeO*~10.5%, TiO_2~1.6%), to ferrobasalts (up to FeO*=13%,TiO_2=3.0%) and tholeiitic andesites or icelandites (SiO_2=58%, TiO_2=2.0%, FeO*/MgO=2.32).

Two other lines of evidence support the importance of fractional crystallization in the origin of the Attu silicic rocks. First, gabbros and intermediate composition plutonic rocks of the albite granite suite commonly have concave-downward REE patterns with positive Eu anomalies suggesting that they are complementary crystal cumulates to the evolved silicic rocks (Figure 12). Second, gabbroic rocks of the albite granite suite are also Fe and Ti-enriched (FeO*=13.3%, TiO_2=4.5%; see sample AT6) and in this way are similar to cumulates that are required by extreme fractionation of tholeiitic basalts [e.g., *Stern*, 1979; *Hunter and Sparks*, 1987]. Note also that the decreased REE and Hf abundances in the most evolved albite granites and rhyolites (SiO_2~77%, Figure 11) suggest that these elements have been removed by fractionation of accessory mineral phases, probably apatite and possibly a small amount of zircon. These minerals are common in rocks of the albite granite suite. While accessory phase saturation is uncommon in modern ocean ridge magmas, it is predicted on experimental grounds and is frequently observed in ophiolites [*DeLong and Chatelain*, 1990].

The isotopic characteristics of the Attu rhyolites suggest that in the course of crystallization they have assimilated some crust. If we assume, on the basis of their REE pattern, that the late dikes are similar to the parental magmas, then low εNd in the silicic differentiates (εNd=8.6-9.7) relative to the late dikes (εNd=9.5-11.5) is consistent with the assimilation of crust (probably marine sediment) during fractionation and emplacement. High $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ in some of the silicic rocks (e.g., AT143) also suggest interaction with marine sediments. Post-emplacement growth may account for some of the high $^{206}Pb/^{204}Pb$ signature if the albite granites have undergone U addition or Pb loss to produce high $^{238}U/^{204}Pb$ either during magma genesis or by hydrothermal alteration [e.g., *Hamelin et al.*, 1984].

Evidence for a Near-Arc Setting and Origin of the Subduction Component

Three geochemical lines of evidence suggest that the Attu sub-arc mantle was variably modified by a subduction

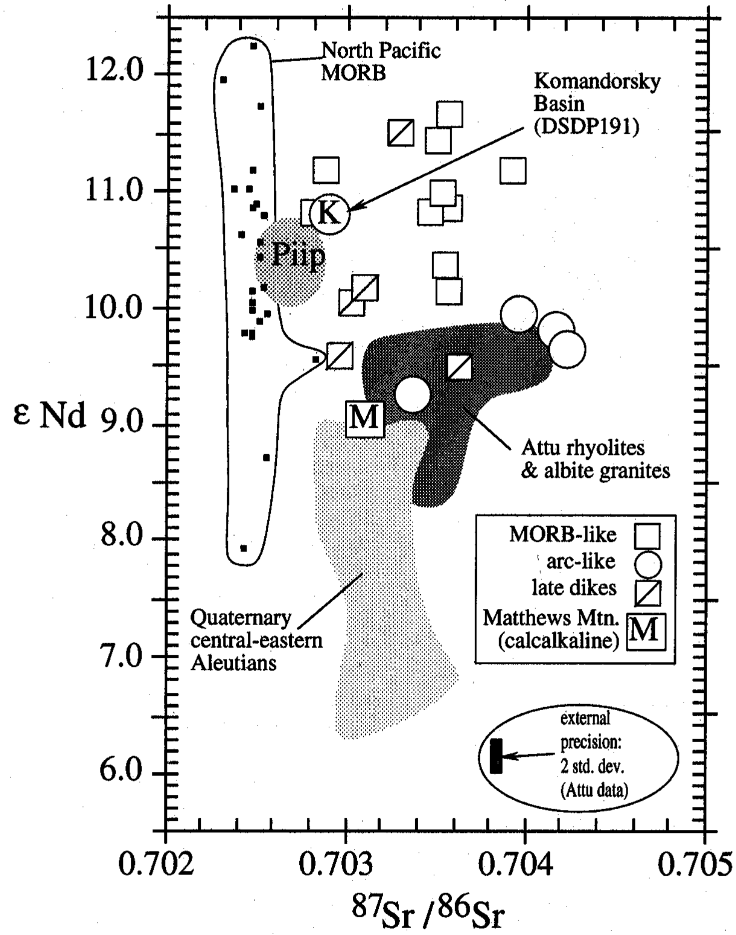

Fig. 8. εNd versus $^{87}Sr/^{86}Sr$. The εNd in the mid-ocean ridge basalt (MORB)-like samples and late dikes overlap (εNd up to ~11.5), but late dikes range to lower values (εNd < 10.0). Arc-like basalts have low εNd (< 10.0) relative to the MORB-like samples. Rhyolites and albite granites (darkly shaded field) have the lowest εNd (~8.6-9.6). Komandorsky Basin basalt and Piip Volcano samples have εNd similar to Attu basement rocks and North Pacific MORB. Attu calcalkaline Matthews Mountain andesite has εNd that straddles the Aleutian and MORB fields. High $^{87}Sr/^{86}Sr$ relative to εNd in the Attu rocks reflects Sr exchange during seawater alteration. Attu data are from Table 3. External precision is based on repeated analyses of standard solutions (see Table 3). North Pacific MORB data (Juan de Fuca and Gorda ridges) are from *White et al.* [1987] and *Hegner and Tatsumoto*, [1987]. Quaternary Central and Eastern Aleutian data are referenced in *Kay and Kay* [1993]. Piip Volcano data are from *Romick et al.* [1993] and *Yogodzinski et al.* [in press]. Komandorsky Basin analysis is from *Kay et al.* [1986].

component. First, relatively high Th/La and La/Ta in the Attu arc-like tholeiites (Figures 5*c* and 6) are broadly similar to those of subduction-related magmas worldwide [e.g., *Pearce*, 1982]. Second, Pb isotope compositions in mafic rocks of the Attu Basement Series form a trend at a high angle to the mantle reference line (Figures 9 and 10), with the highest $^{207}Pb/^{204}Pb$ values occurring in the arc-like samples (i.e., those with high Th/Ta, La/Ta). Third, concentrations of HREE are low relative to compatible elements in the arc-like tholeiites compared to MORB (e.g., Cr-Yb relationship in Figure 7), resulting in a depleted signature that is most common in island arc basalts and most strongly developed in boninites [e.g., *Pearce*, 1982; *Duncan and Green*, 1987].

The chemistry of Attu Basement Series is broadly akin to MORB, but source heterogeneity in the sub-arc mantle is characteristic of subduction-related settings. The arc-like rocks have been found only in southeastern Attu (Figure 2), but there is no geologic basis for separating them from the MORB-like rocks which volumetrically dominate the island; the arc-like and MORB-like tholeiites both occur as pillow lavas and diabase associated with marine sedimentary rocks. The late dikes, which postdate the albite granite suite, are geochemically similar to the MORB-like basalts which dominated the main pulse of mafic magmatism on Attu. Within the Basement Series there was apparently not a succession from MORB-like to arc-like compositions with time. The MORB-like and arc-like magma types appear to have been largely contemporaneous.

Because their source characteristics are chemically akin to those of MORB, the Attu arc-like basalts may provide additional insights into our understanding of the subduction geochemical signature in arc magmas (i.e., the subduction component). Note in particular that ratios among Hf, Ti, and HREE in the Attu arc-type tholeiites are similar to those of MORB but that Ta appears to be depleted relative to other HFSE and HREE (compare sample AT27 and CP98 in Figure 5).

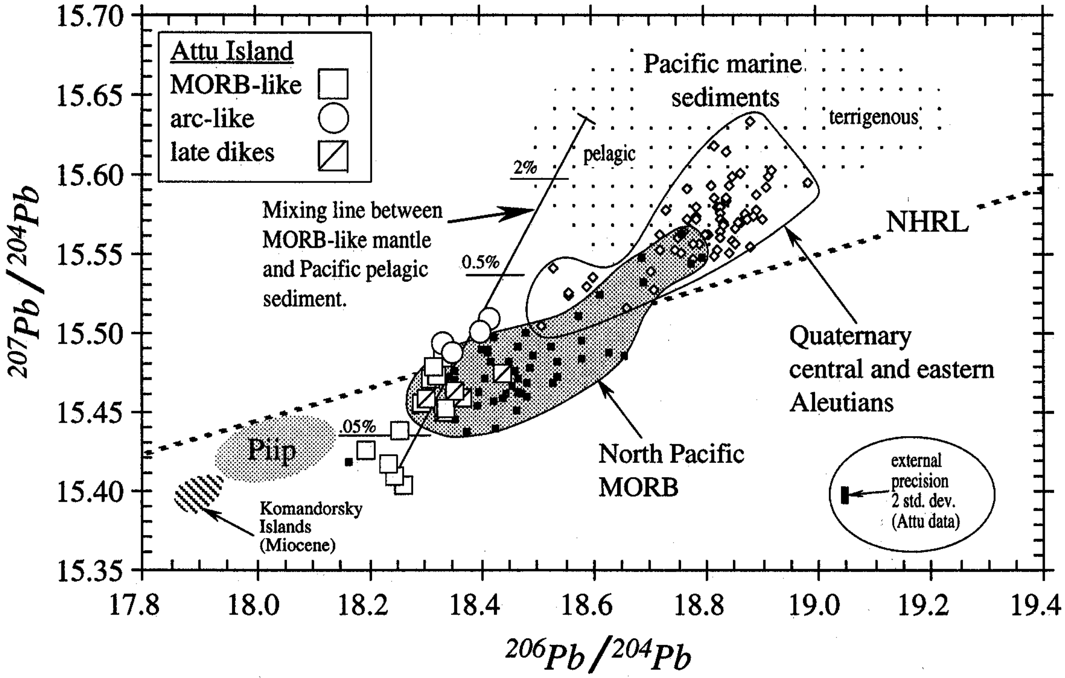

Fig. 9. Pb isotopes. Attu basalts have lower $^{206}Pb/^{204}Pb$ than most North Pacific mid-ocean ridge basalts (MORB) and all Quaternary age volcanic rocks from the central and eastern Aleutians. The mixing line parallel to steep Attu trend is between Pacific pelagic sediment (21.5 ppm Pb, $^{207}Pb/^{204}Pb$=15.63, $^{206}Pb/^{204}Pb$=18.59) and MORB-type Attu mantle (0.08 ppm Pb, $^{207}Pb/^{204}Pb$=15.41, $^{206}Pb/^{204}Pb$=18.25). Piip Volcano Pb values are similar to the least radiogenic Attu basalts. Miocene age calcalkaline rocks of the Komandorsky Islands are the least radiogenic in the Aleutian arc. External precision is based on replicate analyses of standard solutions (see Table 3). Northern hemisphere reference line (NHRL) is from *Hart* [1984]. Field of marine sediments includes ~60 analyses from *Ben Othman et al.* [1989] and *Kay et al.* [1978 and references therein]; North Pacific MORB data are from *Church and Tatsumoto* [1975] and references in Figure 8, Quaternary Aleutian data is referenced by *Kay and Kay* [1993], Piip data are from *Yogodzinski et al.* [1993]. Komandorsky data are from *Housh et al.* [1989], and unpublished Cornell data.

This seems to require that a HFSE-bearing phase has been stabilized in the source of the arc-like magmas. Low Ti contents indicate however, that these magmas are not saturated in such phases [*Ryerson and Watson*, 1987; *Green and Pearson*, 1987]. If residual Ti-rich minerals are to explain the Ta anomalies in these and other arc magmas, then a two-stage process is required wherein fluids or felsic melts, saturated in a Ti-rich mineral (probably rutile), are removed from the subducting slab to contaminate the overlying mantle wedge [e.g., *Ryerson and Watson*, 1987]. *Yogodzinski et al.* [1993] argue that on a regional scale, source enrichment by slab melting is the most reasonable explanation for the arc trace element signature in isotopically MORB-like Western Aleutian volcanic rocks.

The steep trend formed by the Attu data on Pb-Pb plots is also a feature of the subduction component. This trend can be explained by mixing Pb-rich, high $^{207}Pb/^{204}Pb$ marine sediment or sedimentary rock with Pb-poor, low $^{207}Pb/^{204}Pb$ mantle or mantle-derived magma [*Armstrong*, 1971; *Kay et al.*, 1978; *Ben Othman et al.*, 1989]. Such mixing might occur by crust-mantle recycling via subduction, crustal assimilation, or by postmagmatic alteration. Alteration is not an adequate explanation because the Pb isotopic composition of the Attu rocks is correlated with parameters that are largely immune to such processes (e.g., Th/Ta, La/Ta). Crustal assimilation may account for high $^{207}Pb/^{204}Pb$ in some MORB-like samples (evolved, low-Cr MORB-like samples have slightly elevated $^{207}Pb/^{204}Pb$, Tables 1 and 3), but the occurrence of low Ta concentrations (<0.10 ppm) with high $^{207}Pb/^{204}Pb$ in the arc-like tholeiites cannot be produced by assimilation alone. We interpret the Pb isotope signature of the Attu Basement Series rocks as a product of recycling of upper crust into the sub-arc mantle via subduction [*Kay*, 1980]. A simple mixing model between pelagic sediment and depleted peridotite (Figures 9 and 10) shows that a small quantity of Pb from sediment mixed into a Pb-poor MORB-like upper mantle can account for the trend of the Attu Pb data.

Origin of the Matthews Mountain Andesite and the Western Aleutian Calcalkaline Trend

Because calcalkaline volcanic rocks are petrologically complex, simple geochemical models do not in general reproduce their major and trace element characteristics well. Many aspects of modern Central and Eastern Aleutian calcalkaline magmas are nonetheless consistent with an origin by crystal fractionation and magma mixing beginning with parental Mg-rich basalts [*Kay et al.*, 1982; *Kay and Kay*, 1985; *Conrad and Kay*, 1984; *Brophy*, 1990]. These primitive basalts show a range of geochemical characteristics and are known from several locations in the Central and Eastern Aleutian arc [*Kay and Kay*, 1993, Table 1 and Figure 4].

Miocene to Recent calcalkaline rocks from Attu and throughout the Western Aleutians cannot, however, be derived from parental basalts or basaltic andesites like those from the Central and Eastern Aleutians. This is because the Western

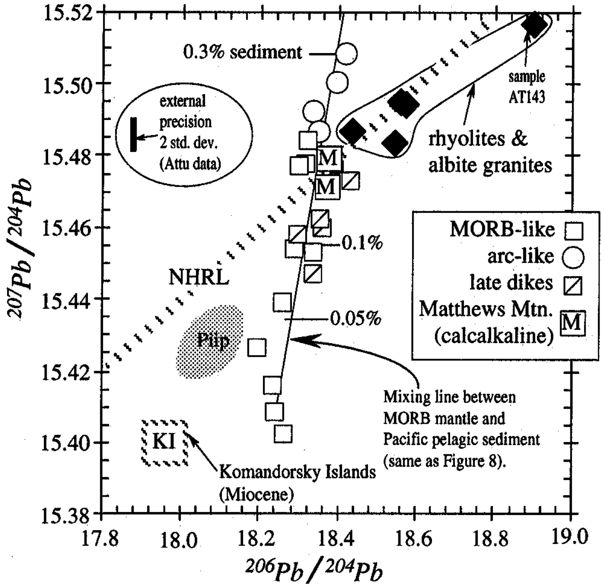

Fig. 10. Pb isotopes in blow-up of Figure 9. Mixing line is as in Figure 9. Note that Matthews Mountain andesites, which have arc-like major and trace element characteristics (e.g., Figures 3 and 6), have Pb isotopes like Attu mid-ocean ridge basalt (MORB)-like rocks. Middle Miocene rocks of the Komandorsky Islands and Piip Volcano have similar characteristics. High $^{206}Pb/^{204}Pb$ in the Attu rhyolites and albite granites may in part reflect in situ radioactive decay if $^{238}U/^{204}Pb$ in these rocks is high relative to the basalts (see text). External precision is based on replicate analyses of standard solutions (see Table 3). Komandorsky Islands values of $^{206}Pb/^{204}Pb$~17.88, $^{207}Pb/^{204}Pb$~15.40 are from *Housh et al.* [1989], and unpublished Cornell data.

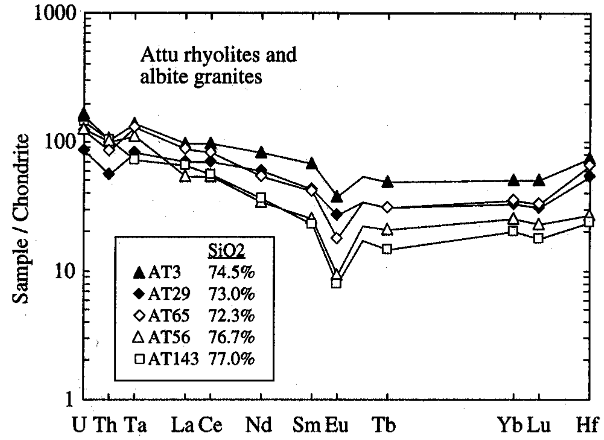

Fig. 11. Chondrite-normalized incompatible elements in Attu rhyolites and albite granites. Note the generally flat rare earth element (REE) patterns and large negative Eu anomalies. Low REE and Hf concentrations in Attu samples AT56 and AT143 (largest Eu anomalies; highest SiO_2) probably result from REE compatibility in accessory mineral phases (e.g., apatite). Note constant Th/U in all samples. Normalizing values are U (0.015), Th (0.050), Ta (0.022), and Hf (0.22). REE normalizing values are as in Figure 4.

Aleutians rocks have generally lower incompatible element concentrations (Figures 13 and 14) and are more strongly calcalkaline (Figure 15) than most Central and Eastern Aleutian volcanic rocks. Model 1 illustrates this point (Figure 17). The parental basaltic andesite in this model was chosen because it has major elements that are relatively close to the daughter (thereby minimizing the amount of fractionation required), and because the shape of its REE pattern is similar to that of the daughter (Figure 17). Closed system fractionation (49%) of an anhydrous mineral assemblage reproduces the major element features of the daughter Matthews Mountain andesite well (Σ residuals^2 = 0.03), but the calculated REE concentrations are too high by more than a factor of 2 (Figure 17). Amphibole fractionation may lower the calculated heavy and middle REE abundances slightly [e.g., *Romick et al.*, 1992], but it is difficult to produce the low CaO/Al_2O_3 of the Western Aleutian rocks without substantial clinopyroxene fractionation (Figures 15 and 17). This is because removal of amphibole (CaO/Al_2O_3 ~0.6 to ~1.5) and/or calcic plagioclase (An85 CaO/Al_2O_3 ~0.55) has little effect on CaO/Al_2O_3 in a primitive basalt (0.60-0.80). Low REE concentrations in the Western Aleutian rocks therefore cannot be the result of extensive amphibole fractionation but must be a characteristic of the parental magma.

The failure of model 1 is not surprising in light of the abundant evidence that Aleutian calcalkaline rocks show for magma mixing [*Kay et al.*, 1982; Conrad and Kay, 1984; *Kay and Kay*, 1985; *Romick et al.*, 1992]. A simple mixture between a primitive Aleutian basalt and an Aleutian dacite is illustrated in model 2 (Figure 17). This kind of mixture reproduces the LREE features of the Matthews Mountain andesite relatively well, but it does not reproduce the HREE features or the low CaO/Al_2O_3 and FeO*/MgO values which characterize the Matthews Mountain andesites and all other Western Aleutian calcalkaline rocks. Choice of different mixing end-members from the Central or Eastern Aleutians will not substantially change this result. We conclude from models 1 and 2 that middle Miocene to Recent calcalkaline rocks of the Western Aleutians cannot be related to Central and Eastern Aleutian basalts or basaltic andesites by reasonable combinations of fractionation and magma mixing.

Model 3 is a fractionation model that begins with a parental magnesian andesite from Piip Volcano (far Western Aleutians, Figure 1). These primitive andesites have low incompatible element abundances (Yb ~5-6 times chondrites) and major element compositions (SiO_2, CaO/Al_2O_3, FeO*/MgO) similar to those of the Matthews Mountain andesites (Figure 17; see also *Yogodzinski et al.* [1993] and *Volynets et al.* [1992]). Model 3 requires only 11% crystallization, and because REE in the parental rock are similar to those in the daughter, the calculated REE also match the observed reasonably well. Model 3 shows that magnesian andesites from Piip Volcano have the general features that are required of parental magmas in the Western Aleutians (low CaO/Al_2O_3, FeO*/MgO, and REE concentrations). This implies that the strongly calcalkaline magmatic rocks from throughout the Western Aleutians (i.e., Figure 15), are more akin to magnesian andesites than to basalts like those of the modern Central and Eastern Aleutian arc. Relatively large major element residuals in model 3 (Σ residuals^2 = 0.38) suggest however, that there is significant variation among Western Aleutian magnesian andesites.

It is important to note that magnesian andesites parental to the Western Aleutian suite are chemically unlike Aleutian "adakites" [*Defant and Drummond*, 1990] which have high Sr (>1500 ppm), high La/Yb (~30), and appear to have originated by melting of the subducting oceanic crust [*Kay*, 1978]. The modeling presented here indicates that Piip Volcano-type magnesian andesites (Sr~380 ppm, La/Yb~4.2) are a reasonable parental composition to the strongly calcalkaline rocks that include graywacke and conglomerate, widespread thinly bedded cherts, cherty argillites, and cherty shales [*Gates et al.*, 1971; *Marlow et al.*, 1973; *Rubenstone*, 1984]. Volcanic rocks within the Basement Series are dominantly mafic with phenocrysts (<10%) of plagioclase, olivine (now replaced by secondary minerals), and clinopyroxene in a generally ophitic to subophitic groundmass. Pervasive hydrothermal metamorphism in these rocks is usually subgreenschist facies but locally reaches amphibolite grade [*Gates et al.*, 1971; *Rubenstone et al.*, 1982; *Vallier et al.*, 1993].

The volcanic and volcaniclastic units are intruded by a series of diabasic sills and dikes and by small plutons of medium-grained isotropic gabbro (diabase-gabbro intrusions of *Gates et al.* [1971]). Figure 2 shows only the largest gabbroic bodies [*e.g., Shelton*, 1986] and the well-developed dike swarm at Fishook Ridge (75% dikes). These rocks range from ophitic diabase to hypidiomorphic gabbro. Mineralogically, they are dominated by plagioclase and clinopyroxene, with some olivine replaced by alteration minerals. There is a subgreenschist metamorphic overprint. Petrographically, the diabasic sills and dikes are broadly akin to the lavas they intrude.

Silicic volcanic and subvolcanic intrusive rocks of the Attu Basement Series occur in small intrusive centers (1-5 km^2 exposures) that form a NE trending belt across Attu Island (Figure 2). These silicic bodies are dominated by albite granite and rhyolite, with only a small amount of intermediate composition plutonic rock. We refer to these rocks collectively as the albite granite suite (Figure 2). The albite granites are medium grained (1-5 mm), generally nonfoliated, compositionally homogeneous, and free of mafic enclaves. They are mineralogically dominated by plagioclase and quartz (> 90 modal % of the rock) with minor clinopyroxene, Fe-Ti oxides, and occasional amphibole (total ~4-10 modal %). Biotite occurs in the mesostasis of the most quartz-rich samples. The rhyolites (including quartz keratophyre of *Gates et al.* [1971]) are generally crystal poor, with phenocrysts of plagioclase, Fe-Ti oxides, clinopyroxene, and occasional amphibole. When present, amphibole phenocrysts are euhedral, nonfibrous, and in some cases zoned, with brown cores and pleochroic blue-green rims. The most silicic rhyolites also contain phenocrysts(?) of rounded and embayed quartz. Intermediate composition rocks of the albite granite suite occur only as small, irregularly shaped plutonic bodies along the (lithologically) heterogeneous margins of the intrusions. These rocks are also mineralogically dominated by plagioclase, but they contain more clinopyroxene and amphibole and less quartz than the albite granites. Chlorite, epidote, and calcite are common alteration minerals, and apatite and zircon are common accessory minerals throughout the albite granite suite. It is an important point that the albite granite suite is volumetrically minor compared to the main mass of mafic rocks that makes up the Attu Basement Series (Figure 2).

The youngest rocks of the Attu Basement Series are a series of dikes, 1-2 m wide, that intrude the rocks of the albite granite suite. We refer to these as the late dikes (Figure 2). They are petrographically similar to the older volcanic rocks and the sills and dikes of the diabase-gabbro intrusive complex, although some (e.g., sample AT21) contain sparse phenocrysts of amphibole and Fe-Ti oxides.

In summary, intrusive rocks of the Attu Basement Series (the diabase-gabbro complex, the albite granite suite, and the late dikes) are thought to be only slightly younger than the volcanic rocks they intrude [*Gates et al.*, 1971]. Analyses presented below indicate that these volcanic and plutonic rocks are related and represent aspects in the development of a single type of magmatic system that characterized early magmatism in the Near Islands.

The Attu Calcalkaline Series

Map units of *Gates et al.* [1971] included within the Attu Calcalkaline Series are the andesitic rocks of Matthews Mountain and the volcaniclastics of the Massacre Bay Formation (Figure 2). These rocks were deposited subaerially on an erosional surface formed on the Basement Series. Radiometric dating of lavas brackets the age of the Attu Calcalkaline Series at 15 to 5 Ma [*DeLong and McDowell*, 1975; *Vallier et al.*, 1993].

Andesitic lavas, dikes, and subvolcanic intrusions at Matthews Mountain (Figure 2) are rich in plagioclase and contain primary hornblende. On neighboring Near Islands (unlabeled in Figure 1), the Attu Calcalkaline Series includes crystal-rich dacite porphyries that contain quartz, plagioclase, amphibole, and biotite [*Gates et al.*, 1971]. These rocks are petrographically and geochemically similar to calcalkaline volcanic rocks of the modern Aleutian arc and thus are readily distinguished from older, mostly mafic rocks of the Attu Basement Series.

GEOCHEMISTRY

Most of the igneous rocks on Attn Island have undergone subgreenschist grade metamorphism. Our interpretation of the Attu data is therefore based largely on elements which are thought to be immobile. These are principally the rare earth elements (REE), Th, and the high field strength elements (HFSE). The consistency of the data presented here provides strong evidence that the magmatic features of the Attu rocks are preserved in these elements. For recent and thorough discussion of element mobility during alteration, the reader is referred to *Dunning et al.* [1991].

Geochemistry of the Mafic Volcanic Rocks

Basalts (SiO_2<53%) and basaltic andesites (SiO_2=53-56%) of the Attu Basement Series have been altered by exchange with seawater during hydrothermal metamorphism; they are Na_2O-enriched (up to 6-7%) and CaO-depleted (down to 6%) with highly variable K_2O (0.20-1.27% ; see Table 1), but their high FeO* contents (up to 12.86%), high FeO*/MgO (up to 2.5), and variably high TiO_2 contents (up to 1.98%) distinguish them as tholeiitic (Figure 3; Table 1; petrographic descriptions of analyzed samples in Table 2). These rocks are mineralogically and texturally uniform but can be grouped into two types, MORB-like and arc-like, on the basis of their trace element characteristics.

The MORB-like tholeiites are relatively depleted in light REE (LREE) and Th, with incompatible trace element characteristics similar to normal MORB (NMORB, Figures 4-6). A few LREE-enriched samples similar to enriched MORB are also included in this group (e.g., HO962A in Figure 5). Concentrations of the heavy REE (HREE; e.g., Yb) relative to compatible elements (Ni and Cr) in these basalts are also generally similar to those in MORB (Figure 7), but some samples have low compatible element concentrations, a characteristic that is commonly seen in arc basalts (Cr less than 30 ppm; see Figure 7). These low-Cr examples nonetheless retain the LREE-depleted signature of MORB and do not show the low HFSE character that is typical of arc volcanic rocks (Figures 5*a* and 6). In general, MORB-like Attu rocks differ consistently from NMORB only in their HREE patterns (note Tb/Yb in Figure 5). The Tb/Yb values of the Attu MORB-like tholeiites average ~ 0.30, which is substantially higher than NMORB Tb/Yb=0.220 of *Sun and McDonough* [1989]. Note, however, that Juan de Fuca MORB also have Tb/Yb greater than Sun and McDonough's NMORB (e.g., Juan de Fuca Tb/Yb ~ 0.255 approximated from *Kay et al.* [1970] and *Wakeman* [1978]). Note also that the HREE

TABLE 3. Isotopic Analyses of Attu and Related Rocks

	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	$^{143}Nd/^{144}Nd$	εNd	$^{87}Sr/^{86}Sr$
MORB-like						
HO962A	18.263	15.402	37.601	0.513165	11.1	0.70291
AT73	18.245	15.408	37.573	0.513150	10.9	0.70360
HO922	18.235	15.416	37.609			
SB1A	18.196	15.426	37.624		11.0*	0.70357*
HO914B*†	18.255†	15.438†	37.624†		10.7*	0.70281*
HO925B*					11.1*	0.70369*
HO925B	18.337	15.452	37.738	0.513179	11.4	0.70353
AT144	18.296	15.454	37.760	0.513191	11.7	0.70359
AT114	18.317	15.478	37.851	0.513112	10.1	0.70359
AT114 (duplicate)	18.309	15.471	37.823	0.513118	10.2	
AT114 (duplicate)	18.322	15.472	37.835	0.513125	10.4	0.70357
AT100	18.330	15.484	37.893	0.513165	11.1	0.70395
AT101	18.313	15.478	37.839	0.513147	10.8	0.70348
Arc-like						
MP8010	18.353	15.486	37.901	0.513104	10.0	0.70397
MP802	18.416	15.508	37.971	0.513089	9.7	0.70424
CP98	18.403	15.500	37.952	0.513097	9.8	0.70418
CP916C	18.336	15.492	37.922	0.513069	9.3	0.70339
Late dikes						
AT27	18.337	15.448	37.810	0.513107	10.0	0.70305
AT16	18.300	15.458	37.756	0.513183	11.5	0.70331
AT50	18.364	15.460	37.861	0.513079	9.5	0.70365
AT22	18.354	15.462	37.834	0.513085	9.6	0.70299
AT21	18.423	15.473	37.904	0.513116	10.2	0.70313
Albite granite suite						
AT3	18.538	15.483	37.958	0.513046	8.8	
AT29	18.582	15.494	38.012	0.513087	9.6	0.70324
AT56	18.571	15.495	37.968	0.513036	8.6	0.70349
AT143	18.901	15.517	38.072	0.513086	9.6	0.70378
AT65	18.435	15.487	37.889	0.513089	9.7	0.70401
Calcalkaline						
AT83	18.379	15.471	37.858			
AT8032	18.372	15.479	37.854		9.3*	0.70306*
Komandorsky Basin						
DSDP-191 basalt*					11.2*	0.70289*

All isotope analyses are by thermal ionization mass spectrometry at Cornell University. Mean standard values and analytical precision (+/- two standard deviations) reported below are based on analyses performed between August and Novermber 1990. Analytical technique for Pb is after *White et al.* [1990]. Measured values for NBS SRM-981 Pb standard were $^{206}Pb/^{204}Pb$=16.907, $^{207}Pb/^{204}Pb$=15.435, $^{208}Pb/^{204}Pb$=36.496. Each ratio was corrected for mass fractionation independently assuming standard values of $^{206}Pb/^{204}Pb$=16.937, $^{207}Pb/^{204}Pb$=15.493, $^{208}Pb/^{204}Pb$=36.705. Analytical precision of +/-0.008 ($^{206}Pb/^{204}Pb$), +/-0.007 ($^{207}Pb/^{204}Pb$), and +/-0.024 ($^{208}Pb/^{204}Pb$) is based on 13 analyses of NBS SRM-981. Nd isotopes by single filament technique after *Walker et al.* [1989]. Ratios were corrected for mass fractionation assuming $^{146}Nd/^{144}Nd$=0.7219. Measured values for LaJolla Nd standard were $^{143}Nd/^{144}Nd$=0.511817 +/-0.000012 and $^{145}Nd/^{144}Nd$=0.348412 +/- 0.000013 based on 15 analyses. Epsilon Nd (εNd) are deviations in 10^4 from present-day chondritic $^{143}Nd/^{144}Nd$ assuming LaJolla εNd= -15.15 [*Lugmair and Carlson*, 1978; *Wasserburg et al.*, 1981]. Sr analyses were done on W single filaments by quadruple collector dynamic procedure. Ratios were corrected for mass fractionation assuming $^{86}Sr/^{88}Sr$=0.11940. Average measured value for NBS SRM-987 Sr standard was $^{87}Sr/^{86}Sr$=0.710228 with analytical precision of +/-0.000042 based on 28 analyses.

*Analysis from *Kay et al.* [1986]; epsilon recalcualted assuming εNd for BRC1=+0.20

†Analyzed at Lamont Doherty by J. L. Rubenstone.

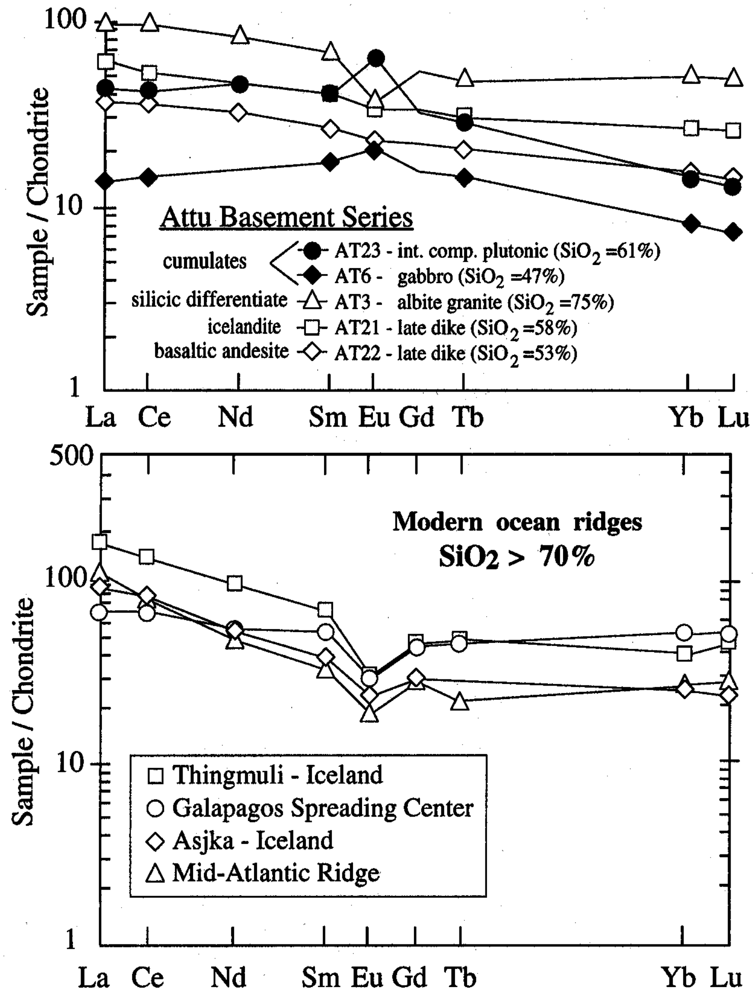

Fig. 12. Comparison of rare earth elements (REE) in cumulate rocks of albite granite suite with late dikes, albite granite, and modern ocean ridge rhyolites. Note the REE patterns in the cumulates are the mirror image of the albite granite (compare La/Sm, Eu/Sm, and Tb/Yb in AT23 and AT3) . Note also the broad similarity of silicic from modern ocean ridges to the Attu albite granite (see also Figure 11). Thingmuli data are from *Wood* [1978], Askja data are from *Shimokawa and Masuda* [1972], Galapagos Spreading Center data are from *Clague et al.* [1981], Mid-Atlantic Ridge data are from *Aldiss* [1981].

Western Aleutian assemblage (i.e., Figure 15 and model 3 in Figure 17). Magnesian andesites of this type are commonly viewed as low-pressure melts of hydrous peridotite [see *Tatsumi*, 1982]. *Yogodzinski et al.* [1993] argue, however, that magnesian andesites of this type may also form through melt-peridotite interaction in the uppermost mantle, immediately beneath calcalkaline volcanoes. This is consistent with the idea that Aleutian calcalkaline volcanism is produced by relatively low temperature fractionation near the base of the arc crust [*Kay et al.*, 1982, *Kay and Kay*, 1985; *Conrad and Kay*, 1984; *Kay and Kay*, 1993; *Brophy*, 1990], and with the observation that Aleutian calcalkaline volcanoes form by relatively small volume eruptions in zones of tectonic compression where magma ascent through the arc crust appears to have been structurally inhibited [*Kay et al.*, 1982; *Geist et al.*, 1988; *Singer and Meyers*, 1992; *Romick et al.*, 1992]. Results from the Western Aleutians imply that fractionation of calcalkaline magmas may begin not within the crust but within the uppermost mantle where the assimilation of relatively hot peridotite is likely [i.e., *Kelemen*, 1986, 1990].

WESTERN ALEUTIAN TECTONICS AND THE ORIGIN OF ATTU AND THE NEAR ISLANDS

Geochemical source variation among the Attu basalts indicates that the Basement Series rocks originated in a near-arc setting. Attu therefore cannot be a tectonically captured slice of Kula or Pacific oceanic crust. The regional trend of decreasing Pb isotope ratios in volcanic rocks from east to west along the Aleutian arc provides direct evidence that the Attu Basement Series rocks formed in situ. In general, $^{206}Pb/^{204}Pb$ values in Miocene to Recent volcanic rocks in the Central and Eastern Aleutians are high (18.5-18.9); in the Western Aleutian Near Islands (Attu) they are low (18.2-18.4); and in the far Western Aleutian Komandorsky region they are very low (17.8-18.2; see Figure 9). Interestingly, this east-west trend is also seen in Pacific ocean ridge and back-arc basin basalts [e.g., *White et al.*, 1987; *Hickey-Vargas*, 1991; *Tatsumoto and Nakamura*, 1991]. Regional source variation of this kind in the Aleutians would be fortuitous indeed, if the volcanic rocks of Attu and the Near Islands were part of an allochthonous terrane.

TABLE 4*a*. Modeling Parameters: Mineral Compositions

	cpx1*	cpx2†	An66††	hbld§	tmtg"	Fo89¶
SiO_2	51.99	51.27	51.68	49.47	0.15	41.43
TiO	0.49	0.40	0.00	0.91	8.51	0.00
Al_2O_3	1.64	3.31	30.40	5.44	6.68	0.00
FeO*	11.04	4.66	0.48	12.18	75.16	10.52
MgO	15.15	16.18	0.00	16.02	3.73	48.14
CaO	18.85	23.09	13.21	10.57	0.00	0.49
Na_2O	0.36	0.20	3.67	1.02	0.00	0.00
K_2O	0.00	0.00	0.11	0.15	0.00	0.00

*Clinopyroxene (SIT5B, [*Kay and Kay*, 1985])
†Clinopyroxene (BUL6B, [*Kay and Kay*, 1985])
††Plagioclase feldspar (75A-113, [*Romick et al.*, 1992])
§Hornblende (75A-110, [*Romick et al.*, 1992])
"Titanomagnetite (75A-21, [*Romick et al.*, 1992])
¶Olivine (4b, [*Conrad and Kay*, 1984])

TABLE 4*b*. Modeling Parameters: Distribution coefficients

	Mineral					Whole-rock	
	hbld*	plag†	cpx††	tmgt§	olv"	SIT-RK5¶	OK4**
La	0.54	0.24	0.10	0.03	0.01	14.61	4.29
Ce	0.84	0.20	0.15	0.03	0.01	32.3	11.03
Nd	1.34	0.14	0.31	0.03	0.01	18.3	6.40
Sm	1.80	0.11	0.50	0.03	0.01	4.68	2.00
Eu	1.56	0.73	0.51	0.03	0.01	1.09	0.70
Tb	2.02	0.06	0.64	0.05	0.01	0.76	0.374
Yb	1.64	0.03	0.62	0.10	0.01	3.42	1.29
Lu	1.56	0.03	0.56	0.12	0.01	0.500	0.199

Some distribution coefficients were determined by interpolation or extrapolatoin. Whole rock values are given in parts per million. Abbreviations are hbld, hornblende; plag, plagioclase; cpx, clinopyroxene; tmgt, titanomagnetite; and olv, olivine.
* *Fujimaki et al.* [1984] average of calcalkaline andesites.
†† *Schnetzler and Philpotts* [1970] andesite GFC271.
¥ *Arth and Hanson* [1975] average for augite in basaltic-to-andesitic rocks.
§Smith and Leeman [1987].
"Values for olivine assumed to be arbitrarily low.
¶Dacite mixing end-member in model 3 [*Romick et al.*, 1992].
**Basalt mixing end-member in model 3 [*Kay and Kay*, 1993].

Vallier et al. [1993] have argued that Western Aleutian islands were formed by arc-type magmatism largely prior to 43 Ma when underthrusting was at a higher angle to the strike of the arc. This implies that growth of the western arc paralleled the early stages in the Central and Eastern Aleutians where voluminous magmatism between 55 and 37 Ma resulted in rapid crustal thickening [*Scholl et al.*, 1970, 1987; *Marlow et al.*, 1973; *Hein and McLean*, 1980]. This view is not, however, consistent with the magmatic development of Attu, because the bimodal, dominantly MORB-like chemistry of the Attu Basement Series rocks is more characteristic of a back-arc basin than an island arc [*Rubenstone et al.*, 1982; *Rubenstone*, 1984; *Kay et al.*, 1986].

Ultimately, the age of the Attu Basement Series will be critical to our understanding of its tectonic origin. Because the magmatic rocks were subjected to seawater metamorphism, and the sedimentary rocks are largely devoid of age-diagnostic fossils, the age of the Attu Basement Series remains poorly constrained. Attempts to date Basement Series basalts by $^{40}Ar/^{39}Ar$ were unsuccessful [*Rubenstone*, 1984]. Metamorphic ages are between 28 and 38 Ma (K-Ar method [e.g., *DeLong and McDowell*, 1975], but if magmatism and metamorphism were largely contemporaneous (see section on evidence for a rift tectonic setting), then some of these dates might approximate emplacement ages. The oldest and most reliable date on a Basement Series rock is 41.5 +/- 2 Ma, obtained on an amphibole separate from a mafic amphibolite [*Vallier et al.*, 1993]. If we assume that tholeiitic magmatism on Attu ended before the beginning of calcalkaline magmatism at ~15 Ma [*DeLong and McDowell*, 1975; *Borusk and Tsvetkov*, 1982;

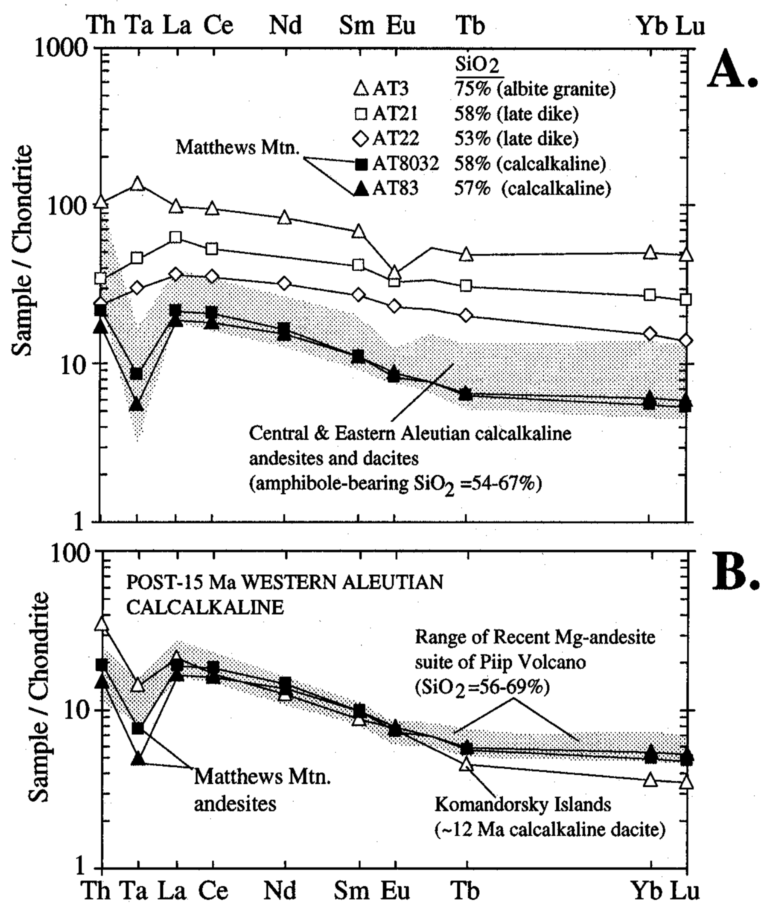

Fig. 13. Chondrite-normalized trace elements of the Attu calcalkaline Matthews Mountain andesites compared to the late dikes and albite granites. In Figure 13*a*, note the low Ta and low overall incompatible trace element concentrations of the Matthews Mountain andesites relative to Basement Series rocks with equivalent SiO_2 (compare AT8032 and AT21). In Figure 13*b*, note the similarity of the Matthews Mountain andesites to the Miocene age calcalkaline dacite of the Komandorsky Islands and magnesian andesite suite from Piip Volcano. Data are from Tables 1 and 2, *Yogodzinski et al.* [1993], and *DeLisio* [1984].

Tsvetkov, 1991], then the Basement Series (including the albite granite suite and late dikes) must be middle Eocene to late Oligocene or early Miocene in age (43? Ma to 25 or 15 Ma). Fossil ages from marine sedimentary rocks of the Near Islands [*Vallier et al.*, 1993] are in accord with this estimate.

A simple tectonic model for the development of the Western Aleutian-Bering Sea region is presented in Figure 18. Rifting and tholeiitic magmatism in the Attu region (Figure 18*a*) are associated with stretching of the overriding plate in response to the switch from oblique to highly oblique convergence following plate reorganization at 43 Ma [*Engebretson et al.*, 1985]. We envision a transtensional regime, broadly analogous to those found in some modern arc systems (e.g., the Philippine Marinduque Basin [*Sarewitz and Lewis*, 1991]). The inferred NE-SW orientation of the rift axis through Attu is consistent with the trend of the Fishhook Ridge dike swarm, the alignment of the silicic intrusive centers (rhyolites, albite granites), and the NE trending belt of well-preserved marine and nonmarine sedimentary rocks that overlie the volcanic-dominated basement (Nevidiskov, Chuniksak, and Faneto formations, Figure 2). These sedimentary rocks were probably deposited within a rift depression prior to and during uplift of the Attu Basement Series. Local amphibolite grade metamorphism and occurrence of a 400-m-thick sill complex in the central part of the island [*Vallier et al.*, 1993] are consistent with a rift interpretation. We assume that clockwise rotation of Attu in dextral Western Aleutian shear zones has been relatively minor; this is consistent with the findings of *Geist et al.* [1988] which suggest that only 3° of rotation of the Near Island block has occurred in late Cenozoic time. Paleomagnetic data from Attu are inconsistent and do not provide a reasonable constraint on rotation of the Near Island crustal block (W. Harbert, personal communication, 1992).

Note that the NE-SW orientation of the rift axis and the presence of EW-oriented (not NS-oriented) magnetic lineations in the Bowers Basin immediately north of the Near Islands [*Cooper et al.*, 1992, Figure 3] argue against formation of the Attu Basement Series by simple back-arc spreading behind the Bowers island arc. We assume that the Bowers subduction system largely predates 43 Ma (Figure 18*a*). Subduction beneath the Bowers Ridge and spreading in the Bowers Basin may however, have continued through the Middle Tertiary [*Cooper et al.*, 1992, figure 7]. If so, then the Attu Basement Series may simply be an extension of Bowers Basin crust onto the Western Aleutian Ridge crest [e.g., *Kay et al.*, 1986, Figure 3b]. Figure 18*a* is based on the general idea that after plate reorganization at 43 Ma [*Engebretson et al.*, 1985], the

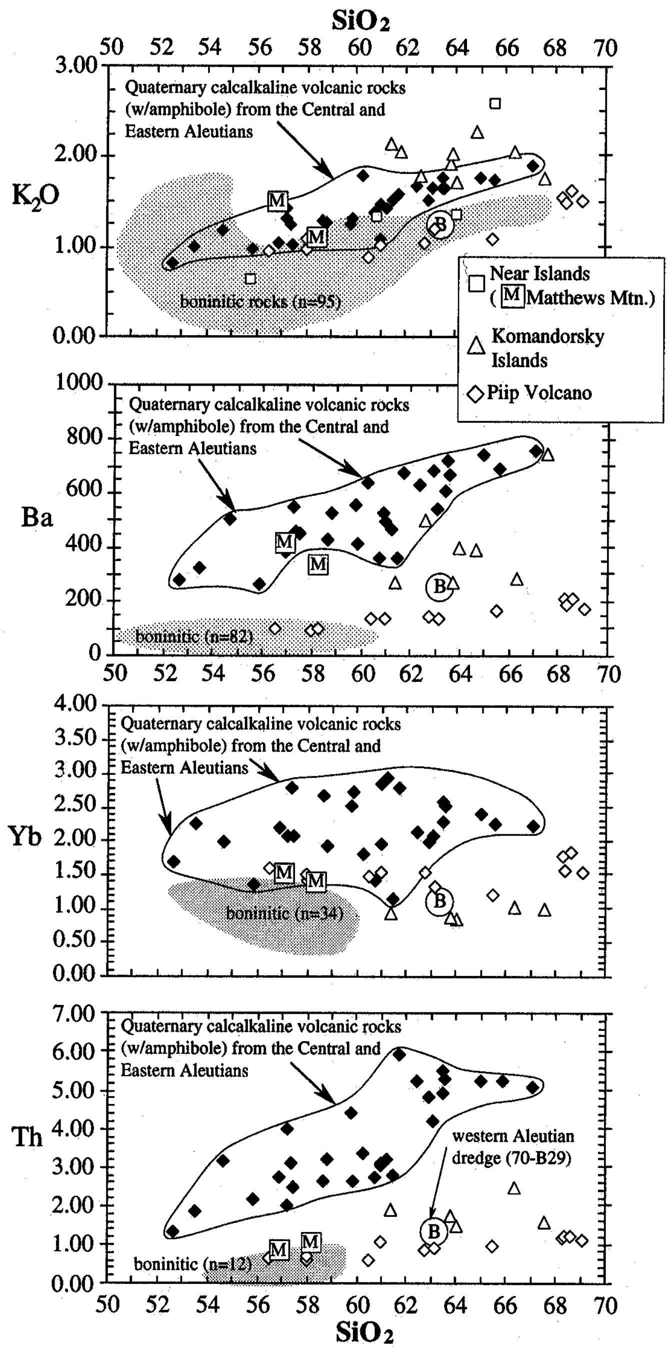

Fig. 14. Silica variation diagrams for post-15 Ma Western Aleutian calcalkaline rocks compared with central and eastern Aleutian calcalkaline rocks (solid symbols) and with boninites and related rocks (gray fields). Note that Ba, Yb, and Th contents in the western arc are lower than those in the Aleutian reference suite, but that K_2O contents broadly overlap. Boninite fields provide comparison with arc-related magmas with low abundances of incompatible elements. Western Aleutian dredge sample labeled "B" (Quaternary age) was collected between Attu and Buldir islands [*Scholl et al.*, 1976] (complete analysis by *Kay and Kay* [1993]). Western Aleutian data sources are the same as in Figure 15.

tectonics of the Western Aleutian Ridge proper must have been dominated by strike-slip faulting (see also *Cooper et al.* [1992, Figure 7D]).

Baranov et al. [1991] have shown that the trenchward (western) margin of the Shirshov arc was rifted to the west during the Neogene opening of the Komandorsky Basin, while the eastern margin of the Shirshov arc remained approximately stationary. This is shown in Figure 18*b* as a simple arc-splitting [*Karig*, 1971], and in this respect it differs from the transform-style rifting mechanism of *Baranov et al.* [1991].

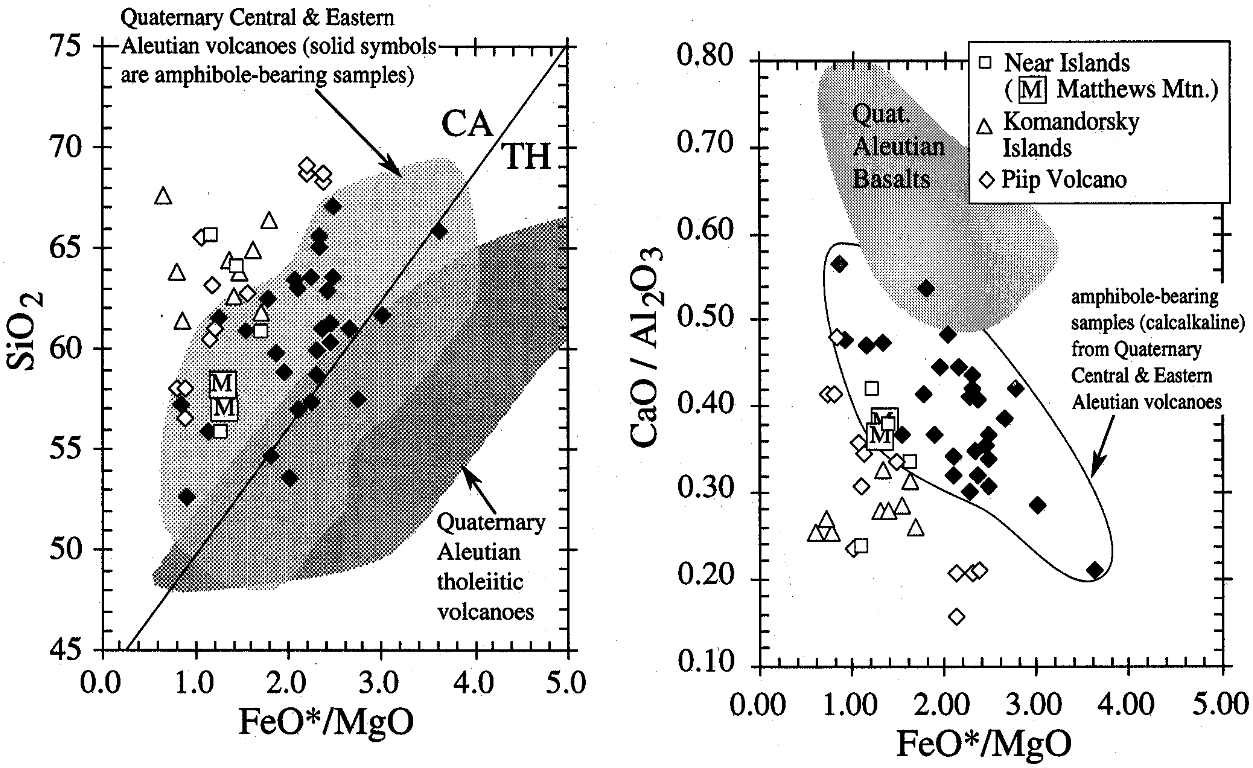

Fig. 15. Post-15 Ma Western Aleutian calcalkaline rocks compared to central and eastern Aleutian basalts and calcalkaline suite. Note low FeO*/MgO relative to SiO_2, and low CaO/Al_2O_3 relative to FeO*/MgO in the Western Aleutian suite making them more strongly calcalkaline than the central and eastern Aleutian assemblage. Data are from *Gates et al.* [1971], *Borusk and Tsvetkov* [1982], *DeLisio* [1984], *Kay and Kay* [1993], and *Yogodzinski et al.* 1993].

In any case, active rifting was presumably accompanied by voluminous tholeiitic magmatism which formed the Neogene oceanic-type basement that underlies the Komandorsky Basin [*Cooper et al.*, 1987].

The eruption of calcalkaline volcanic rocks in the Near and Komandorsky islands beginning 15 m.y. ago, marks a profound change in the tectonic style of the Western Aleutian arc at that time. Chemically, the calcalkaline rocks of the Near and Komandorsky islands are analogous to late Pleistocene rocks of Piip Volcano. Piip Volcano is centered on small dilational structures within the broadly transpressional tectonic regime of the modern Western Aleutian arc [*Baranov et al.*, 1991; *Seliverstov et al.*, 1990*a*, *b*; *Geist et al.*, 1988]. We therefore infer that the onset of calcalkaline magmatism 15 m.y. ago, marks the switch from transtensional to transpressional tectonics in the western arc. This is consistent with the observation that calcalkaline, subduction-related magmatism did not become active along the full length of the modern Aleutian arc until approximately middle Miocene time [*Kay et al.*, 1990].

The model in Figure 18 does not account for the topographically elevated Komandorsky Islands in the westernmost Aleutians (the "Komandor block" of *Seliverstov et al.* [1984]). The crust of the Komandorsky Islands was probably thickened by an Early to Middle Tertiary episode of subduction-related magmatism [e.g., *Borusk and Tsvetkov*, 1982; *Tsvetkov*, 1991], but such an episode is difficult to reconcile with contemporaneous rift-related magmatism in the Near Islands. One possibility is that the Komandorsky block is largely allochthonous, having been moved by strike-slip motion along the Western Aleutian Ridge to its present location in the westernmost arc [*Scholl et al.*, 1987; *Geist et al.*, 1991; *Cooper et al.*, 1992]. Westward motion of the Komandorsky block may have occurred in conjunction with the opening of the Komandorsky Basin and rifting of the (former) Shirshov arc (e.g., Figure 18*b*). This kind of model allows for the distinctly different early magmatic histories (>30 Ma) of the Komandorsky and Near Island groups, and it suggests that the alkaline, arc-type magmatism in the Komandorsky Islands at 20-25 Ma [*Borusk and Tsvetkov*, 1982; *DeLisio*, 1984; *Housh et al.*, 1989; *Tsvetkov*, 1991] may be associated with unusual tectonic conditions at the Shirshov-Aleutian junction immediately prior to or during arc splitting and westward transport of the Komandorsky block (see Figure 18*b*).

The onset of calcalkaline magmatism in the Western Aleutians may have resulted from the clogging of the Aleutian-Kamchatka junction by the collision of the westward drifting Komandorsky block at ~15 Ma (Figure 18*c*, see also *Scholl et al.* [1987] and *Geist et al.* [1991]). In this interpretation, the Komandorsky Islands would have arrived at their present location prior to the youngest Komandorsky magmatic episode (middle to late Miocene [*Borusk and Tsvetkov*, 1982; *Tsvetkov*, 1991]. There are no Pleistocene or Recent magmatic rocks on any of the Western Aleutian Islands, but seafloor imaging (GLORIA images, A. Stevenson, personal communication, 1990) has shown that there is a zone of active magmatism along the northern margin of much of the Western Aleutian Ridge. Piip Volcano, which is located not on the Aleutian Ridge proper but on back-arc crust of the southernmost Komandorsky Basin (Figure 1), is probably the westernmost extent of this active zone of Western Aleutian

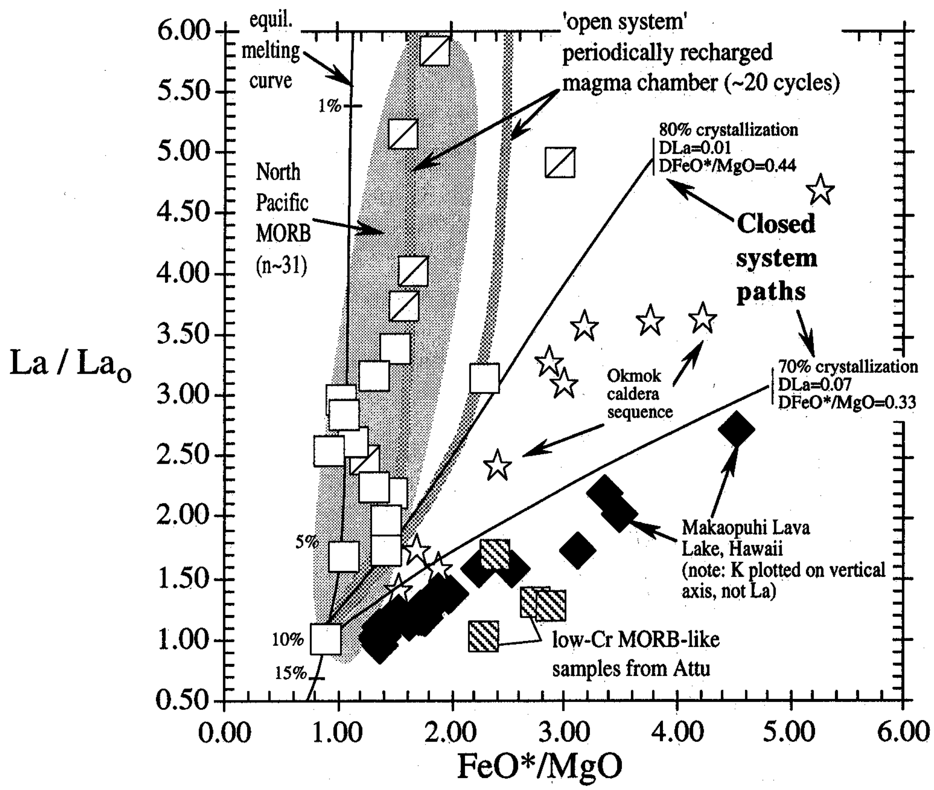

Fig. 16. Attu basalts compared to modeled crystallization and melting trends. Parental composition in crystallization models is the least evolved Attu Basement Series basalt (sample HO925B, Table 1*a* - La=2.33, FeO*/MgO=0.89). La/La$_0$ (enrichment factor) is La concentration in the residual melt divided by the La concentration in the parental magma. Closed system models are simple Rayleigh fractionation. Steep trend of most Attu data requires low *D*La and high *D*FeO*/MgO (*D*=mineral/melt bulk distribution coefficient). Open system model ('periodically recharged magma chamber" [*O'Hara*, 1977]) assumes a constant magma volume with parameters chosen to maximize the steepness of the trend (38% crystallization, 2% eruption, 40% recharge, DLa=0.01, DFeO*/MgO=0.44). High FeO*/MgO line in the open system model are expected values prior to recharge and mixing; low FeO*/MgO line is the residual magma after recharge and mixing. Data fitting this "open system" model would fall between these lines. Note that the Attu data fall between the open system model and the equilibrium melting line. Equilibrium batch melting model [*Shaw*, 1970, equation 11) is adjusted to pass through the parental Attu basalt at approximately 10% melting (*D*La=0.01, *D*FeO*/MgO=0.27). Makaopuhi lava lake data from *Wright and Okamura* [1977] show the trend of closed system fractionation in a natural system. Data from Eastern Aleutian Okmok caldera (location in Figure 1) are from *Kay and Kay* [1993 and references therein]. North Pacific MORB data are from *Wakeman* [1978]. Figure is modified from *Kay et al.* [1982].

magmatism. The northward shift of the Western Aleutian magmatic front between late Miocene and Pleistocene time probably came at 5 Ma with the 5° northward shift of Pacific plate motion relative to North America [*Cox and Engebretson* 1985].

Alternatives to the tectonic model in Figure 18 are many. All models, however, will remain speculative until the ages of regionally important structures (e.g., Bowers and Shirshov ridges) become more clearly known (see also *Cooper et al.* [1992]). We emphasize primarily that the early tectonic events in the Western Aleutians (43-20 Ma) as more closely allied to contemporaneous activity in the Western Bering Sea than to the development of the central and eastern arc; the chemistry of the Attu Basement Series rocks seem to require this. We believe further that the transition to calcalkaline magmatism on Attu and throughout the Western Aleutian islands in the middle Miocene marks a change to a compressional tectonic regime at that time. The predominance of calcalkaline magmatism in compressional regimes and tholeiitic magmatism in extensional regimes has been noted previously in the Aleutians [e.g., *Kay et al.*, 1982; *Singer* **and Myers**, 1992]. We emphasize that these associations are particularly well developed in the history of Attu and western part of the arc (see also *Yogodzinski et al.* [1993]).

CONCLUSIONS

The Attu Basement Series is most similar to back-arc basin crust, but the age and structure of the Attu Basement Series and its position on the crest of the modern Aleutian Ridge do not readily provide for its formation within the framework of arc-parallel rifting (e.g., the Mariana system). Figure 18 provides a model for another style of arc-adjacent, oceanic rift, wherein the lithosphere of the overriding plate is stretched in response to tectonic adjustments resulting from a switch from convergence to strike-slip-dominated tectonics. On Attu, magmatism in this setting produced oceanic-type crust with the geologic, metamorphic, and geochemical signature of a

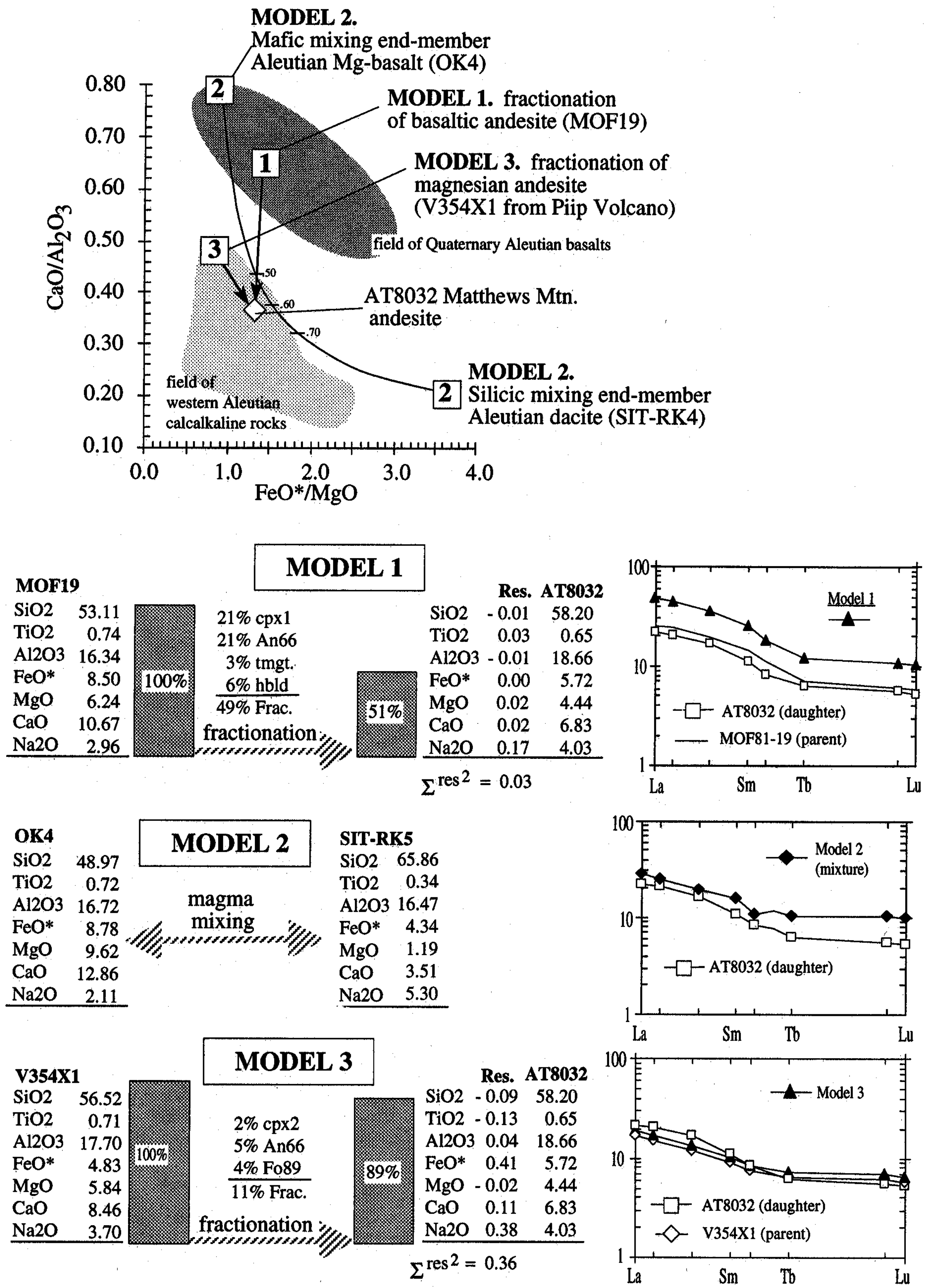

Fig. 17. Simple crystal fractionation and magma mixing models for the origin of the calcalkaline Matthews Mountain andesite (AT8032) demonstrate the distinction between calcalkaline rocks of the Western versus Central and Eastern Aleutian arc. In model 1, major elements are modeled by least squares to obtain mineral percentages removed in Rayleigh fractionation models for the rare earth elements (REE). Good fit to the major elements in model 1 requires 49% crystallization, but this produces calculated REE concentrations that are too high (compare REE in AT8032 and model 1). Note that removal of amphibole with the aim of dampening REE

subduction-related ophiolite (e.g., suprasubduction zone ophiolite of *Pearce et al.* [1984]). The Attu Basement Series is, however, not allochthonous. Rather, it formed in situ by rift-related magmatism within the broadly transtensional tectonic setting that was established in the Western Aleutians after plate reorganization at ~43 Ma. The back-arc type crust of the Attu Basement Series was eventually uplifted, more or less in place, in the transpressional regime of the Late Tertiary Western Aleutian Ridge [e.g., *Kay et al.*, 1986, Figure 3*b*].

Diffuse generation of oceanic crust and the association of rift tectonics with major strike-slip boundaries provide a mechanism for the in situ formation and emplacement of suprasubduction zone ophiolites [*Sarewitz and Lewis*, 1991]. The Andaman Sea, Cayman Trough, Gulf of California, and the Philippine Marinduque Basin provide actualistic models for the formation of the Western Aleutian arc crust. While the scales and the tectonic geometries associated with these examples vary widely, they illustrate the kinds of processes that may have operated in the formation and preservation of the Attu Basement Series. These kinds of processes are increasingly recognized in the geologic record. Examples may include the Troodos or other Tethyan ophiolites [e.g., *Moores et al.*, 1984], the Rocas Verde ophiolite of southern Chile [*Dalziel*, 1981], and the Andaman ophiolite of northern Sumatra [*Ray et al.*, 1988].

The Attu Basement Series also has important consequences for arc magma formation in the modern Aleutian system. The history of oblique subduction and rift-related magmatism in the Western Aleutian arc has provided an opportunity to sample Aleutian sub-arc mantle that was relatively unmodified by subducted components. The geochemistry of the tholeiitic basalts of the Attu Basement Series requires that in Early to Middle Tertiary time, the Aleutian sub-arc mantle at ~176°E was like that of MORB. This is important, because in the more conventional tectonic geometry of the Central and Eastern Aleutian arc, recycling, mixing, and melting are complete to the extent that even in the most primitive magmas, it is difficult to know whether the unmodified mantle source was more similar to MORB or OIB [e.g., *Kay*, 1980; *Morris and Hart*, 1983; *Perfit and Kay*, 1986]. The Attu Basement Series tholeiites presumably escaped contamination by recycled upper crust because recycling was minimized by oblique subduction, and because eruption rates in the rift environment were relatively high (see also *Yogodzinski et al.* [1993]).

The eruption of a strongly calcalkaline series throughout the Western Aleutians beginning approximately 15 m.y. ago, heralded the end of transtensional tectonics and the beginning Western Aleutian style transpressional tectonics and associated magmatism. Late Pleistocene activity at Piip Volcano is a continuation of this magmatic episode. At Piip Volcano, hydrous (and oxidized?), amphibole-bearing melts of intermediate-to-silicic composition (mostly 60-68% SiO_2) have leaked to the surface along a small zone of tectonic dilation within the broadly transpressional Western Aleutian Ridge [see *Seliverstov et al.*, 1990*a*, *b*; *Baranov et al.*, 1991; *Romick et al.*, 1993; *Yogodzinski et al.*, 1993]). We infer from geochemistry and regional magamtic history that the modern transpressional tectonic style of the Western Aleutian arc [*Geist et al.*, 1988; 1991] was established approximately 15 m.y. ago.

Miocene to Recent magmas that have erupted in the Western Aleutians cannot be produced by crystal fractionation of basalts or basaltic andesites like those found in the central and eastern arc. The Western Aleutian calcalkaline trend is produced by the removal of amphibole and Fe-Ti oxides from primitive magnesian andesites [see *Yogodzinski et al.*, 1993]. This implies that calcalkaline magmatism is tied not only to specific crystallization conditions within the arc crust [e.g., *Kay and Kay*, 1985], but also to distinctive conditions within the sub-arc mantle conducive to the formation of primitive andesites. These may include shallow and hydrous melting conditions [e.g., *Tatsumi*, 1982], small percentage melting, and/or an extended melt-peridotite reaction history (i.e., mantle assimilation [*Kelemen*, 1986, 1990]).

The role of magnesian andesites in the Central and Eastern Aleutians remains unclear. Primitive mafic xenocrysts (Mg number >88) in andesites and dacites of some of the most strongly calcalkaline Aleutian centers (e.g., Buldir Volcano [*Kay and Kay*, 1985]) has been taken as good evidence that primitive basalts [*Kay et al.*, 1982; *Nye and Reid*, 1986] are parental to the calcalkaline trend in the Central Aleutians. The recognition of magnesian andesites at Piip Volcano provides another possible source for primitive xenocrysts in calcalkaline lavas.

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increases during fractionation [e.g. *Romick et al.*, 1992], will produce large major element residuals, because amphibole removal from basalts or basaltic andesites cannot produce the low CaO/Al_2O_3 values seen in the Western Aleutian rocks (see text). Model 2 is a mixing model between primitive Aleutian Mg-basalt (OK4) and a dacite from the Central Aleutians (SIT-RK4). Note that high Al_2O_3 in the Matthews Mountain andesite cannot be explained by mixing the end-members in model 2. Similarly, the low CaO/Al_2O_3 relative to FeO*/MgO of the Western Aleutian suite cannot be explained by such a mixing line or by any other mixing line using end-members selected from the central or eastern Aleutian arc. Most importantly, no combination of fractionation and mixing of magmas from the central and eastern Aleutians (models 1 and 2) can explain the Western Aleutian data (compare with Figure 15). In model 3, simple fractionation of a Piip Volcano magnesian andesite reproduces the major elements reasonably well with only 11% crystallization. This small degree of crystallization also results in a relatively good match for the REE. Modeling parameters are given in Table 3.

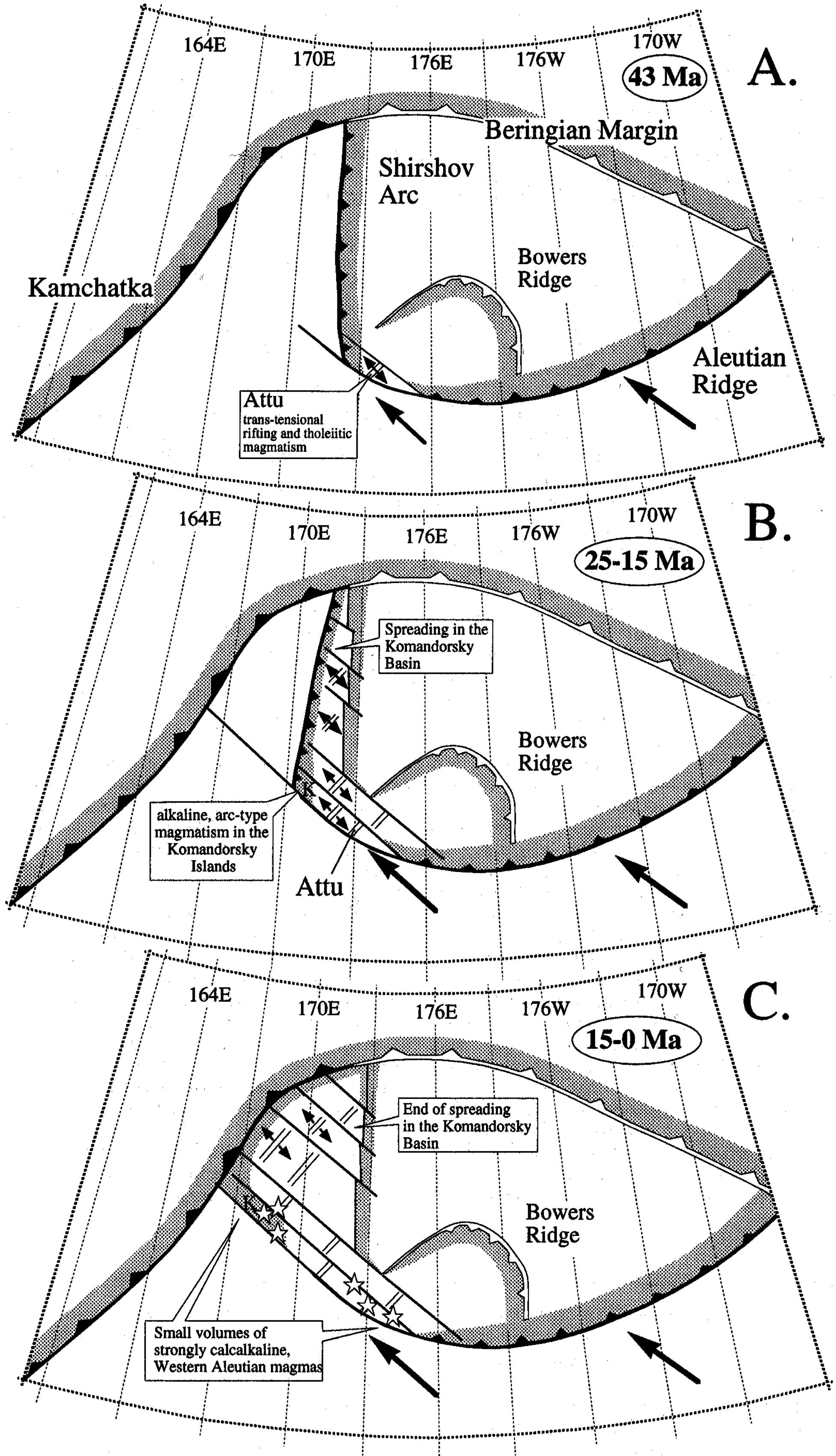

Fig. 18. Tectonic cartoon for the origin of the Near Islands and the Western Aleutian-Bering Sea region. (*a*) At ~43 Ma, change to highly oblique convergence in the Western Aleutians results in transtensional rifting and tholeiitic magmatism at the Aleutian-Shirshov junction to form the Attu Basement Series. The aseismic Shirshov Ridge (Figure 1) is shown as the northwestern extension of the pre-43 Ma Aleutian island arc [*Scholl*

et al., 1989; *Baranov et al.*, 1991; *Cooper et al.*, 1992]. This tectonic geometry was probably accommodated by convergence across the Bering Sea [*Scholl et al.*, 1989; *Cooper et al.*, 1992]. The Bowers Ridge is assumed here to have become inactive by ~43 Ma. This is a simplifying assumption as the Bowers arc may have been active into the middle Tertiary (see discussion in text and *Cooper et al.* [1992]). (*b*) At approximately 25 Ma, breakup of the Shirshov island arc and westward opening of the Komandorsky Basin occur in conjunction with the westward migration of arc-related crustal blocks (terranes) that will eventually become accreted to the Kamchatka margin and possibly form portions of the Komandorsky Islands in the westernmost Aleutians (Figure 1). Komandorsky alkaline volcanic rocks with arc-type geochemical signatures are approximately 20-25 m.y. old [*Borusk and Tsvetkov*, 1982; *Tsvetkov*, 1991]. (*c*) Collision of arc-related terranes with the Kamchatka trench changes the tectonic regime of the Western Aleutians from transtensional to transpressional (see also *Watson and Fujita* [1985], and *Geist et al.* [1991]). This results in the observed shift from voluminous tholeiitic magmatism to volumetrically minor, strongly calcalkaline magmatism in the Near Islands (Matthews Mountain andesites) and in the Komandorsky Islands beginning ~15 m.y. ago. This kind of magmatism continues along the northern margin of the Western Aleutian Ridge today (e.g., Piip Volcano). Stars are locations of post-15 Ma calcalkaline magmatism along the Western Aleutian Ridge.

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