Geochemical evolution of topaz rhyolites from the Thomas Range and Spor Mountain, Utah

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Abstract

Two sequences of rhyolite lava flows and associated pyroclastic deposits are exposed in the Thomas Range (6 m.y.) and at Spor Mountain (21 m.y.) in west-central Utah. Both contain topaz indicative of their F-enrichment (>0.2%) and aluminous nature. The rhyolites are part of the bimodal sequence of basalt and rhyolite typical of the region. Moderate changes in major elements coupled with large variations in trace elements in vitrophyres from the Thomas Range are generally consistent with fractionation of observed phenocrysts. Especially important roles for the trace minerals are suggested. Nonetheless, disagreement between major and trace element models regarding the degree of crystallization as well as improbably high $\overline{D}^{Ta}$ and $\overline{D}^{Th}$ suggest that some diffusive differentiation involving the migration of trace elements complexed with volatiles also occurred *or* that monazite, titanite or samarskite fractionation was important. Higher concentrations of F in the evolving Spor Mountain rhyolite drove residual melts to less silicic compositions with higher Na and Al and promoted extended differentiation yielding rhyolites extremely enriched in Be, Rb, Cs, U, Th, and other lithophile elements at moderate SiO_2 concentrations (74%).

Introduction

During the Late Cenozoic much of the western United States experienced eruptions of bimodal suites of basalt and rhyolite (Christiansen and Lipman, 1972) consequent to the development of the Basin and Range province, the Snake River Plain, and the Rio Grande rift. These rocks are especially characteristic of the Great Basin where bimodal rhyolitic and mafic volcanism commenced about 22 m.y. ago (e.g., Best et al., 1980). Those rhyolites which are rich in fluorine (up to 1.5 wt.%) and lithophile elements (Be, Li, U, Rb, Mo, Sn, and others) appear to be part of a distinct assemblage of topaz-bearing rhyolite lavas that occurs across much of the western United States and Mexico. The general characteristics of topaz rhyolites have been discussed by Christiansen et al. (1980, 1983a,b). Perhaps the best known are those from the area near Topaz Mountain in west-central Utah (Fig. 1).

Two age groups of topaz rhyolite lava flows and their associated tuffs occur in this region, one forming the Thomas Range and the other peripheral to Spor Mountain (Fig. 1). The rhyolites of the Thomas Range have been mapped as the Topaz Mountain Rhyolite (Lindsey, 1982.

1979). The lavas and tuffs were emplaced 6 to 7 m.y. ago from at least 12 eruptive centers (Lindsey, 1979) and coalesced to form a partially dissected plateau with an area in excess of 150 km². Turley and Nash (1980) have estimated their volume to be about 50 km³.

The topaz rhyolite of the Spor Mountain Formation has an isotopic age of 21.3±0.2 m.y. (Lindsey, 1982). This rhyolite and its subjacent mineralized tuff (the "beryllium tuff") occur in isolated and tilted fault blocks on the southwest side of Spor Mountain and as remnants of domes and lava flows on the east side of the mountain. Lindsey (1982) suggests that at least three vents were active during the emplacement of the rhyolite of Spor Mountain. Eruptive episodes for both the Spor Mountain and Thomas Range sequences commenced with the emplacement of a series of pyroclastic breccia, flow, surge, and air-fall units and were terminated by the effusion of rhyolite lavas (Bikun, 1980).

This study documents the petrologic and geochemical relationships of these lavas. Particular attention is paid to the role of fluorine in the evolution of these rocks and to the origin of their extreme enrichment in incompatible trace elements. We conclude that most of the variation is the result of the fractionation of the observed phenocrysts; "thermogravitational diffusion" appears to have played only a small role.

The petrology of these rhyolites was initially studied as part of an investigation of uranium deposits associated with fluroine-rich volcanic rocks (Burt and Sheridan,
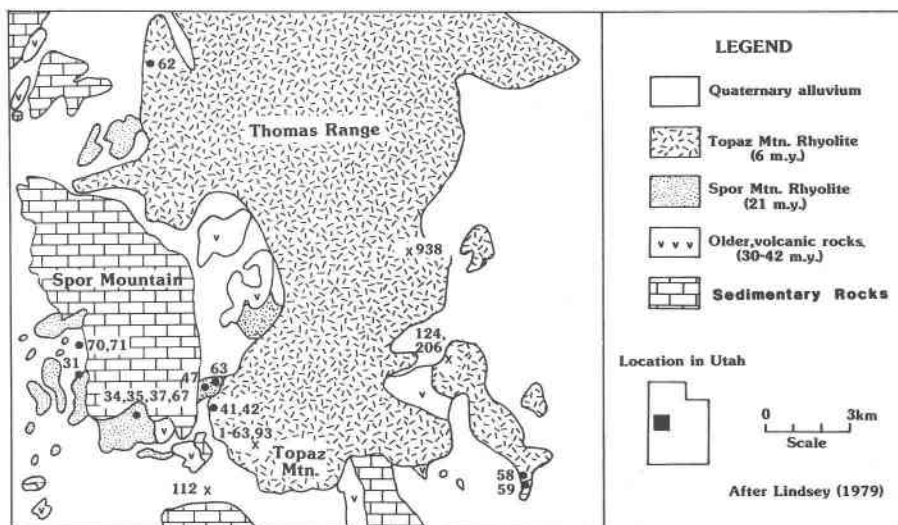
Fig. 1. Generalized geologic map of the southern Thomas Range, Utah (after Lindsey, 1979, 1982; Staatz and Carr, 1964). Sample localities shown by dots (surface) and crosses (drill core). Sample numbers correspond to those in Tables 1 and 2. Samples 61a, b, and c were collected from a site north of the area shown and sample 6-358 and 18-113 are from a drill core obtained from a site just to the west of the map area (Morrison, 1980).

1981). The rhyolite of Spor Mountain is underlain by a cogenetic tuff which is mineralized with Be, F, and U (Lindsey, 1977, 1982; Bikun, 1980) and is associated with sedimentary uranium deposits and small fluorite deposits. Presumably the lavas or their intrusive equivalents were the source of the ores. In contrast, the Topaz Mountain Rhyolite is not associated with economic surface mineralization, but is similar in many respects to the older lavas. There is also increasing evidence that such fluorine-rich volcanic rocks are intimately associated with subvolcanic porphyry molybdenum deposits (Burt et al., 1982).

Analytical methods

The samples analyzed in this study were selected from a suite of more than 100 samples collected from outcrops in addition to drill-core samples obtained by Bendix Field Engineering Corporation (Morrison, 1980). Samples selected for chemical analysis were chosen to represent the spectrum of fresh rock compositions on the basis of petrographic examination.

Wavelength dispersive XRF was used to determine Si, Ti, Al, Fe, Mn, Ca, K and P. Analyses were performed using glass discs (Norrish and Hutton, 1969). Energy dispersive XRF with undiluted rock powders was used to determine Rb, Sr, Y, Zr, Nb and Ga. Atomic absorption spectrometry was used to analyze for Na, Mg, Be, Li and Sn. Natural rock powders were used as standards in each case. Fluorine was analyzed using ion-selective electrodes and a few samples were also analyzed for F and Cl using an ion chromatograph as described by Evans et al. (1981). Estimates of analytical precision are included in

the tables. Other trace elements were analyzed by instrumental neutron activation at the University of Oregon and are considered to be precise to better than ±5% except for Cr and Eu (±10%).

Mineralogy

The mineralogy of the two sequences of rhyolites are grossly similar. The Topaz Mountain Rhyolite generally contains sanidine, quartz, sodic plagioclase±biotite±Fe-rich hornblende±titanite. Biotite compositions show increasing Fe/(Fe + Mg) with differentiation (Turley and Nash, 1980). Two-feldspar and Fe–Ti oxide temperatures correlate well and range from 630° to about 850°C (Turley and Nash, 1980; Bikun, 1980). Augite is found in one lava flow (SM-61) with a high equilibration temperature. Spessartine–almandine garnet occurs in some lavas, generally as a product of vapor-phase crystallization (Christiansen et al., 1980). Titanomagnetite is present as microphenocrysts in most samples, but ilmenite is rare.

The rhyolites of the Spor Mountain Formation contain sanidine, quartz, sodic plagioclase, and biotite. Biotites are extremely Fe-rich (Christiansen et al., 1980) and suggest crystallization near the QFM oxygen buffer. Two-feldspar temperatures cluster around 680° to 690° at 1 bar. Magnetite is more abundant than ilmenite.

Magmatic accessory phases in both rhyolites include apatite, fluorite, zircon, and allanite. Topaz, sanidine, quartz, Fe–Mn–Ti oxides, cassiterite, garnet, and beryl occur along fractures, in cavities and within the groundmass of the rhyolites. They are the result of crystallization from a vapor released from the lavas during cooling and devitrification.

Rock chemistry

Major element composition

Chemical analyses of vitrophyres and fresh felsites from both volcanic sequences are presented in Tables 1 and 2, along with CIPW normative corundum or diopside as calculated on a water- and fluorine-free basis (Fe^{2+}/(Fe total) = 0.52; Bikun, 1980). All of the samples have high Si, Fe/Mg, and alkalies and low Ti, Mg, Ca, and P, features shared with all topaz rhyolites and typical of rhyolites from bimodal (mafic-silicic) associations from around the world (Ewart, 1979; Christiansen et al., 1983a). As expected, both suites are generally high in fluorine (greater than 0.2%). Only two vitrophyres, one from each sequence of lavas, are peraluminous (corundum normative). Devitrification of the vitrophyres may result in compositions with normative corundum, for example compare SM-61a, b, and c or samples SM-34 and 35 which were collected from different "facies" of the same lava flows. In both cases the intensely devitrified samples are corundum normative. This is apparently related to the loss of an alkali-bearing vapor-phase during their subaerial crystallization (cf. Lipman et al., 1969).

Vitrophyres from the two rhyolites can be distinguished by their major element geochemistry. In general, the rhyolites from the Thomas Range have greater concentrations of Si, Mg, and Ti, and less Al and F than the older rhyolites from the Spor Mountain Formation.

Whole-rock compositions are plotted in terms of normative quartz-albite-orthoclase in Figure 2 for comparison with experimental work in the same system (with H_2O or with H_2O–F). Normative femic constituents are generally less than 2% of the total, so when contoured in terms of An content, this systems is probably a reasonable model for the magmas under consideration. Although the vitrophyres from the Spor Mountain Formation show considerable scatter, there is no overlap with those from the Thomas Range. Both suites lie on the feldspathic side of the hydrous minima. Only sample SM-35, which displays an anomalously high calcium content, lies in the primary phase field of quartz on this diagram. With decreasing Ca, Mg, Ti, and increasing F, the samples from the Thomas Range plot nearer the fluid-saturated minimum. We interpret this trend (also noted by Hildreth, 1977) as a reflection of lower temperatures and the approach to fluid-saturation of an initially unsaturated magma crystallizing at about 1 to 1.5 kbar total pressure. The displacement of the Spor Mountain rhyolites toward the Ab apex is probably the result of their higher fluorine content. Manning (1981) has shown that increasing F concentration shifts the fluid-saturated minimum at 1 kbar to increasingly albitic compositions (Fig. 2). A poorly constrained path defined by residual melts in a cooling granite–HF–H_2O system shows this same trend

Table 1. Whole-rock chemical analyses of rhyolites from Spor Mountain, Utah

Lithology* Sample No.	V 31	F 34	V 35	F 37	F 47	F 63	M 67	V 70	V 71	F 6-358	F 18-113	F 26-112	****
weight per cent													
SiO_2	73.3	74.8	73.9	75.3	74.5	73.7	65.4	72.2	73.7	74.0	74.4	73.5	1
TiO_2	0.06	0.06	0.06	0.07	0.05	0.04	0.94	0.04	0.06	0.06	0.06	0.03	3
Al_2O_3	13.2	13.3	13.1	12.7	14.4	13.8	14.6	13.9	13.3	13.5	13.4	14.0	2
Fe_2O_3	1.48	1.52	1.43	1.01	0.81	1.17	6.35	1.19	1.33	1.63	1.48	1.19	5
MnO	0.07	0.04	0.06	0.08	0.02	0.05	0.09	0.06	0.05	0.07	0.06	0.06	5
MgO	0.05	0.07	0.08	0.07	0.08	0.21	0.36	0.06	0.07	0.23	0.11	0.14	10
CaO	0.61	0.60	1.27	0.70	0.14	0.45	2.87	0.58	0.70	0.43	0.59	0.63	4
Na_2O	4.02	3.76	4.33	3.85	2.63	3.83	3.15	4.90	4.39	3.92	4.01	3.94	2
K_2O	4.99	4.81	3.65	4.86	5.95	4.91	4.95	4.98	4.74	4.88	4.83	4.85	2
P_2O_5	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.00	20
Total	97.78	98.96	97.88	98.64	98.58	98.16	98.71	97.90	98.35	98.72	98.94	98.34	
LOI**	2.96	0.50	3.55	2.57	1.43	0.91	2.30	3.00	2.98	0.77	0.57	0.70	10
parts per million													
Rb	970	1000	1060	605	1600	1400	576		1000	860	1030	1450	4
Sr	5	10	5	5	5	20	345		30	40	10	15	10
Y	92	135	135	46	13	56	145		122	94	140	77	6
Zr	99	120	120	120	85	80	360		110	105	120	87	6
Nb	90	135	120	73	130	120	45		114	105	130	130	4
Ga	50	50	n.d.	40	70	65	30		45	35	75	70	15
CIPW normative minerals ***													
C	0.07	0.81	–	–	3.40	1.40	–	–	–	1.01	0.47	1.11	
di	–	–	0.64	0.37	–	–	1.80	1.42	0.66	–	–	–	

* V= vitrophyre, F= felsite, T= tuff, S= spherulitic vitrophyre.
** LOI= Loss on ignition at 900°C for four hours.
*** Calculated from volatile – free analyses.
**** Estimated precision based on replicate analyses. Reported as one standard deviation expressed as percentage of amount reported.
n.d.= not determined.

Table 2. Whole-rock chemical analyses of rhyolites from the Thomas Range, Utah

Lithology Sample No.	V 42	T 43	F 58	V 59	V 61a	F 1-93	F 8-938	S 29-124	V 29-206	S 61b	F 61c	V 62a	F 62b	S 1-63
SiO_2	75.9	75.9	75.8	76.6	75.0	75.9	75.8	75.9	76.2	74.2	74.2	76.1	75.4	75.5
TiO_2	0.09	0.16	0.08	0.07	0.28	0.07	0.08	0.07	0.07	0.28	0.30	0.14	0.14	0.10
Al_2O_3	12.9	12.7	12.9	12.8	13.1	12.8	12.6	13.0	12.7	13.1	13.8	12.4	12.5	12.7
Fe_2O_3	1.09	1.17	0.99	0.98	1.26	1.61	0.93	1.00	0.99	1.25	1.35	1.04	1.00	1:02
MnO	0.08	0.06	0.05	0.06	0.04	0.08	0.07	0.07	0.07	0.06	0.06	0.06	0.05	0.06
MgO	0.09	0.13	0.17	0.06	0.19	0.13	0.13	0.18	0.05	0.26	0.21	0.12	0.30	0.14
CaO	0.74	0.74	0.99	0.67	0.82	0.39	0.69	0.72	0.72	1.06	0.85	0.76	1.97	0.46
Na_2O	4.11	3.45	4.12	3.98	3.48	3.76	3.92	3.65	3.74	3.43	3.52	3.35	3.58	4.11
K_2O	4.69	5.02	4.82	4.62	5.51	4.83	4.72	4.71	4.80	5.46	5.56	5.40	5.15	5.02
P_2O_5	0.00	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.02	0.01	0.00	0.00	0.00
Total	99.99	99.33	99.92	99.84	99.70	99.57	98.94	99.30	99.34	00.12	99.86	99.37	100.09	99.11
LOI	3.11	2.66	0.79	3.34	2.97	0.33	0.34	2.16	2.84	2.80	0.17	2.83	1.66	0.16
parts per million														
Rb	585	374	580	533	186	614	440	470	433		185	377	372	515
Sr	12	19	31	15	83	6	24	5	2.5		79	25	59	11
Y	43	47	61	93	35	33	30	80	75		37	42	37	24
Zr	115	140	105	110	190	115	90	105	100		205	130	110	130
Nb	70	50	60	60	40	72	51	48	47		40	50	42	61
Ga	30	40	45	35	30	45	45	40			35	45	20	30
CIPW Normative Minerals														
C	–	–	–	0.02	–	0.65	–	0.57	–	–	0.49	–	–	–
di	0.75	0.34	1.70	–	0.03	–	0.44	–	0.07	0.15	–	0.70	2.22	0.66

See footnotes to Table 1.

(Kovalenko, 1977b). These considerations suggest that 1.5 kbar may be a maximum estimate for the crystallization pressure.

Trace element composition

The trace element compositions of these rhyolites (Tables 1, 2, and 3) are especially informative; notable are the substantial enrichment of Rb, Cs, Nb, Y, U, Th, Ta, and HREE and the depletion of Eu, Sr and Co. These features distinguish topaz rhyolites from most other rhyolitic magma types. The highly fractionated character of these rhyolites is demonstrated by their similarily to Li-pegmatites and their associated granites (e.g., Goad and Cerny, 1981), and to ongonites (topaz-bearing volcanic
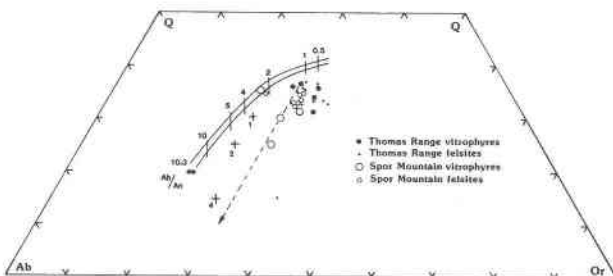

Fig. 2. Normative composition of rhyolites in terms of quartz (Q), albite (Ab) and orthoclase (Or) compared to experimentally determined ternary minima in the hydrous system (contours from Anderson and Cullers, 1978) and in a hydrous fluorine-bearing system (Manning, 1981). Numbers on solid contours indicate PH_2O and beside crosses they indicate weight % F in water-saturated system at 1 kbar. The dashed line is from Kovalenko (1977) and represents the composition of residual melts formed by fractional crystallization in a hydrous granite system with 1 to 2 wt.% HF at 1 kbar and 800 to 550 °C.

rocks from central Asia, Kovalenko and Kovalenko, 1976).

Topaz Mountain Rhyolite. Rhyolites from the Thomas Range display considerable variation in their trace element composition even though they are all fairly enriched in lithophile (or "fluorophile") trace elements. Significantly, vitrophyres with low equilibration temperatures, like SM-29-206 (630°C), are more enriched in lithophile elements than those with high-equilibration temperatures. With increasing evolution (as indexed by uranium concentration) F, Be, Sn, Li, Rb, Cs, Na, Th, Ta, Nb, Y and the HREE increase whereas K, Sr, Zr, Hf, Sc, Mg, Ca, Ti, P, and the LREE plus Eu are depleted. REE patterns thus become flatter (lower La/Lu) and Eu anomalies deeper (lower Eu/Eu*) with increasing evolution in the Topaz Mountain Rhyolite (Fig. 3). Variation diagrams for some of these elements are shown in Figure 4. Enrichment factors for two rhyolites are shown in Figure 5.

The correlation of these changes with the eruptive history is the subject of work in progress, but the coherent variation of many trace and major elements across the suite (plus their mineralogic and temporal similarity) suggests that the rhyolites are part of a comagmatic group of rocks.

Spor Mountain Formation. The samples analyzed from the Spor Mountain Formation are more enriched in U, Th, Rb, Cs, Be, Li, Sn, Ta, Ga, Nb, Y and REE (especially HREE) than the younger rhyolites from the Thomas Range. The rhyolites of the Spor Mountain Formation have extremely low K/Rb (30 to 44), high Rb/Sr (ca. 150), and low Mg/Li (2 to 5). Semiquantitative analyses reported by Lindsey (1982) are compatible with these results. Thus the composition suggests that the Spor Mountain magmas were highly differentiated. The

Table 3. Trace element composition of topaz rhyolites from Spor Mountain and the Thomas Range, Utah

| | THOMAS RANGE | | | | | | SPOR MTN. | | | |
	SM-61a	SM-62a	SM-29-206	SM-42	SM-41	SM-31	SM-35	SM-70	SM-71	**
Li	20	40	60	–	–	–	80	100	60	10
Cs	2.5	7.4	9.7	14.1	14.2	55	58	–	–	2
Be	2	6	7	–	–	–	52	82	65	10
Cr	n.d	n.d	2.6	n.d	1.4	3.6	3.2	–	–	10
Co	0.4	n.d	n.d	n.d	n.d	0.4	0.4	–	–	5
Sc	2.8	1.8	1.7	2.2	2.4	2.6	2.7	–	–	3
Sn	2	–	–	–	–	–	–	30	–	15
Hf	7.9	5.5	4.9	5.7	6.0	6.2	7.1	–	–	·5
Ta	2.3	4.5	5.4	6.4	6.0	26	25	–	–	2
Th	25	60	47	56	65	64	69	45	67	3
U	5	15	19	25	27	36	38	26	31	4
F	1900	1600	3200	5400	6400	8200	11000	12500	8000	5
Cl	–	–	–	–	–	–	1680	1060	–	5
La	73	–	18	25	27	60	59	–	–	2
Ce	132	129	53.1	64	62	144	137	–	–	3
Nd	63	35	31	31	35	52	51	–	–	3
Sm	8.8	–	7.5	4.9	5.0	13.5	17.8	–	–	2
Eu	1.2	0.21	0.065	0.059	0.051	0.15	0.044	–	–	7
Tb	1.3	1.2	1.9	–	0.79	3.3	3.4	–	–	4
Yb	4.0	9.8	9.6	7.9	8.8	15.6	15.7	–	–	3
Lu	0.53	0.90	1.7	1.6	1.5	2.5	2.6	–	–	3
La/Lu$_N$	14.0	–	1.1	1.6	1.8	2.4	2.3	–	–	–
La/Ce$_N$	1.4	–	0.88	1.05	1.12	1.07	1.11	–	–	
Eu/Eu*	0.35	0.06	0.02	0.02	0.025	0.024	0.006	–	–	–

Note: All analyses in ppm; ratios relative to chondrites.
** Estimated precision, reported as one standard deviation as percentage of amount present.
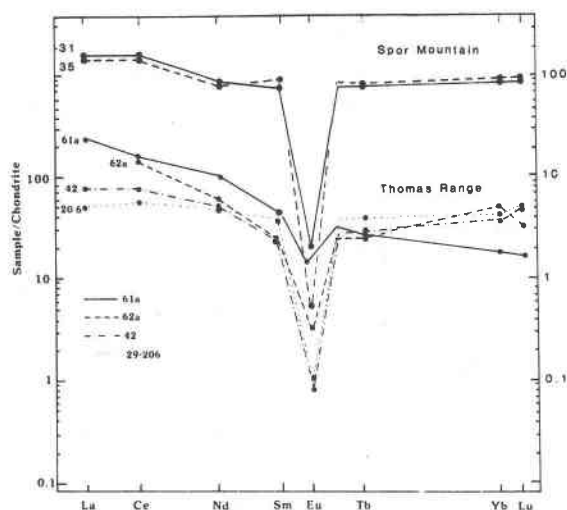n.d. = not detected.

Fig. 3. Rare-earth element composition of the rhyolites from the Thomas Range and Spor Mountain. Note the enrichment of HREE and the depletion of LREE with increasing depth of the Eu anomaly in samples from the Thomas Range. The scale on the right applies to samples from Spor Mountain and the scale on the left to samples from the Thomas Range. Concentrations are normalized to 0.83 times Leedey chondrite (Masuda et al., 1973).
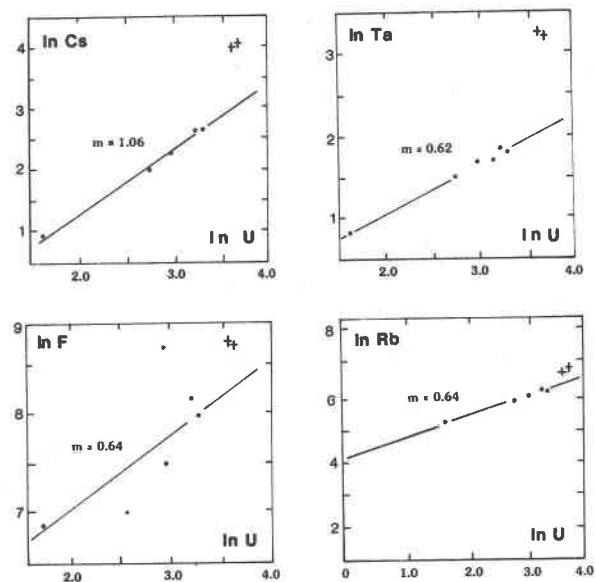

Fig. 4. Logarithmic variation diagrams for topaz rhyolites from the Thomas Range (·) and Spor Mountain (+). The lines represent least-squares fits to the data from the Thomas Range. The bulk partition coefficient ($\overline{D}$) for each element is derived from the equation $\overline{D} = 1 - m$, where m is the slope of the line (see text).
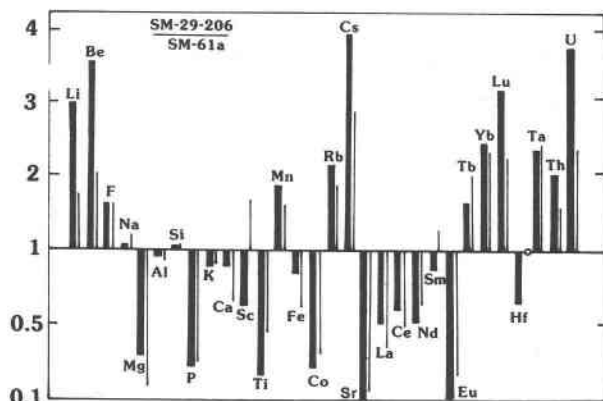
Fig. 5. Enrichment factors for 30 elements derived by comparing two samples from the Thomas Range (heavy bars). Elemental ratios derived by comparing early and late parts of the eruption of the Bishop Tuff (Hildreth, 1979) are shown for comparison (fine bars).

anomalous concentrations of lithophile elements are consistent with the suggestion that the Be–U–F ores in the underlying beryllium tuff were formed as these elements were released and mobilized by devitrification and leaching of the lavas (Bikun, 1980; Burt and Sheridan, 1981). Mineralization is absent from the less evolved rhyolites of the Thomas Range.

The REE patterns (Fig. 3) also provide evidence of extreme differentiation in terms of the large negative Eu anomalies (Eu/Eu* = 0.03 to 0.01). In addition the patterns are nearly flat (low La/Lu$_N$ = 2.5). Similar patterns are recognized in other highly evolved rhyolitic magmas (Hildreth, 1979; Keith, 1980; Bacon et al., 1981; Ewart et al., 1977; and others), and are typical of many topaz rhyolites (Christiansen et al., 1983a).

Petrologic evolution

The origin of the elemental variations and substantial enrichments of incompatible trace elements (those which have bulk partition coefficients $\overline{D}$ less than 1) in topaz rhyolites could result from: (1) small and variable degrees of partial melting, (2) crystal fractionation, or (3) liquid fractionation (e.g., thermogravitational diffusion, liquid immiscibility, or vapor-phase transport). In this section evidence is presented supporting the suggestion that the chemical evolution of the Topaz Mountain Rhyolite (and by analogy the rhyolite of Spor Mountain as well) was governed by crystal fractionation possibly acting in concert with some form of liquid differentiation.

Topaz Mountain Rhyolite

Partial melting. Variable degrees of partial melting of a uniform source could possibly produce the variation observed in the incompatible trace elements without substantially altering the major element chemistry of the liquid (as long as no phase was eliminated from the

source). The simplest discriminant between partial melting and fractional crystallization is the behavior of compatible trace elements (elements with bulk partition coefficients $\overline{D}$ greater than 1). In feldspar-rich rocks, Eu, Ba, and Sr all have $\overline{D}$ greater than 1. Large ranges in the concentrations of these elements cannot be produced by partial melting processes as the range is limited by $1/\overline{D}$ (Shaw, 1970; Hanson, 1978).

Figure 6 shows the behavior of Eu relative to U (with a very low $\overline{D}$). In quantitative formulations, the fraction of liquid produced by partial melting or the residual liquid from crystallization (both symbolized by f) are inversely related to uranium concentration, if uranium behaves as an incompatible element. As noted below, $\overline{D}^u$ is slightly greater than 0; nonetheless the trend of the data is clearly inconsistent with batch partial melting, but indicates a $\overline{D}^{Eu}$ of about 3 for some type of fractionation process.

There appears to be little evidence of partial melting processes remaining in these highly evolved rhyolites. A determination of the partial melting history would require a detailed examination of a number of less silicic consanguineous rocks (like SM-61a) or perhaps an examination of their intrusive equivalents.

Fractional crystallization. Many of the elemental variations such as increasing Na, U, Th, and Rb, and decreasing Ca, Mg, Ti, and Eu with increasing SiO$_2$ content of the rhyolites of the Thomas Range are qualitatively consistent with fractional crystallization of observed phenocrysts. The major and trace element data combine to suggest that samples SM-42, -62, and -29-206 could be derived from SM-61 by crystal separation. This process was modeled using a program written by J. Holloway and
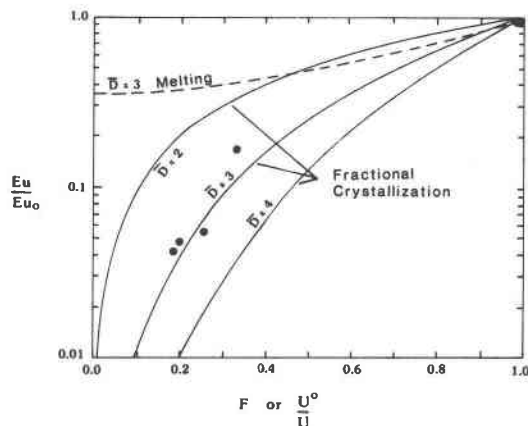

Fig. 6. Behavior of compatible (Eu) versus incompatible (U) elements during batch partial melting and fractional crystallization. Curves were calculated from

$$\frac{C}{C_0} = \frac{1}{\overline{D}(1 - f) + f} \text{ (batch partial melting) and } \frac{C}{C_0} = f^{(\overline{D}-1)}$$

(fractional crystallization). The Eu–U variation does not appear to be related to variable degrees of partial melting.

M. Hillier that conforms in most respects with suggestions of Reid et al (1973). Weighting is based on 1/wt.% of each component, normalized so that SiO_2 has a weight of 1. The major element compositions of the rocks (except P_2O_5) and phenocrysts were used as input for the program. Feldspar (Ab/An/Or), magnetite, ilmenite and biotite (Fe/Mg) compositions were varied within limits determined by microprobe analyses of phenocrysts (Bikun, 1980; Turley and Nash, 1980). Table 4 shows the results of this attempt and demonstrates that fractional crystallization of quartz, sanidine, plagioclase, biotite, and Fe–Ti oxides (magnetite and ilmenite in a 50:50 mixture) can account for most of the chemical variation. Mn and Mg show the largest (but statistically insignificant) relative errors. The SiO_2 content of the "cumulates" ranges from 72 to 74%. These models suggest that from 40 to 60% crystallization of a magma similar to SM-61 could produce residual liquids similar to the other three samples. The relative proportions of the phenocryst phases are in accord with those generally observed in the vitrophyres with sanidine > quartz > oligoclase > biotite > Fe–Ti oxides. Although apatite was not included in the fractionating mineral assemblage (because of constraints imposed by the program), mass balance relations suggest that removal of 0.0006 to 0.0012 wt. fraction apatite could account for the excess P_2O_5 in the calculated compositions. Although these models are in no way unique, they do suggest that substantial fractional crystallization could have occurred without drastically changing the major element (Si, Al, K, Na) composition of the derivatives—a qualitative argument usually proffered to discount fractional crystallization of granitic magmas to produce substantial enrichments of incompatible elements. Although Hildreth (1979) has made a strong case for the lack of crystal *settling* in the magma chamber of the Bishop Tuff, there is considerable evidence (field, theoretical, and experimental) that many granitic intrusions solidify inwards by crystallization on the walls and especially the floor—without significant crystal settling (Bateman and Chappell, 1979; Bateman and Nokleberg, 1978; McCarthy and Fripp, 1980; Groves and McCarthy, 1978; Pitcher, 1979; McBirney, 1980). Theoretically, the magma chamber that gave rise to these topaz rhyolites could have crystallized substantially between eruptions of the lavas. In fact, the elevated concentrations of fluorine would promote extensive crystal fractionation. Wyllie (1979) has suggested that the addition of fluorine to a granitic magma results in a substantial reduction of the solidus temperature but produces a smaller reduction of the liquidus temperature. This would extend the duration of fractional crystallization of a cooling pluton over a wider temperature (and time) interval. The accumulation of fluorine in the residual melt would further enhance this effect. In addition, fluorine may reduce the viscosity of silicate melts and increase diffusion rates (Bailey, 1977) which would aid crystallization and fractionation.

A more rigorous test (at least one with fewer assumptions) of the crystal fractionation model can be made by using the trace element compositions and the analytical method of Allegre et al. (1977). Using the conventional Rayleigh fractionation law and trace element variation diagrams that use elements with very low bulk partition coefficients as "monitors" of fractionation, U and Cs in this case, the bulk partition coefficient for a fractionating system can be derived without further assumptions as outlined below. This bulk partition coefficient can then be used to determine f (the degree of fractional crystallization) and to place some limits on the composition of the fractionating mineral assemblage.

During perfect fractional crystallization the Rayleigh fractionation law (Neumann et al., 1954) can be used to describe the trace element concentration of a differentiated liquid, C_L, relative to the concentration in the parent melt, C_0:

$$C_L = C_0 f^{(\overline{D}-1)} \qquad (1)$$

where $\overline{D}$ is the bulk partition coefficient. If $\overline{D} \ll 1$ (incompatible in the mineral assemblage, denoted by *),

Table 4. Least-squares approximations of the effect of fractional crystallization of rhyolites from the Thomas Range

	observed derivative (%)	calculated derivative (%)	fractionating phases (wt. %)	
a. SM-61a to SM-62a				
SiO_2	76.1	76.1	sanidine	16.55
TiO_2	0.14	0.13	quartz	10.00
Al_2O_3	12.4	12.4	plagioclase	8.66
Fe_2O_3	1.04	1.04	biotite	1.21
MnO	0.06	0.05	Fe-Ti oxides	0.47
MgO	0.12	0.11		
CaO	0.76	0.76	Residual liquid	63.19
Na_2O	3.35	3.35		
K_2O	5.40	5.40		
Total	99.37	99.35		
b. SM-61a to SM-29-206				
SiO_2	76.2	76.2		
TiO_2	0.07	0.07	sanidine	22.29
Al_2O_3	12.7	12.7	quartz	11.96
Fe_2O_3	0.99	0.99	plagioclase	6.06
MnO	0.07	0.05	biotite	1.28
MgO	0.05	0.07	Fe-Ti oxides	0.59
CaO	0.72	0.84		
Na_2O	3.74	3.76	Residual liquid	57.73
K_2O	4.80	4.81		
Total	99.34	99.50		
c. SM-61a to SM-42				
SiO_2	75.9	75.9		
TiO_2	0.09	0.07		
Al_2O_3	12.9	12.9	sanidine	29.53
Fe_2O_3	1.09	1.09	quartz	18.69
MnO	0.08	0.06	plagioclase	9.21
MgO	0.09	0.06	biotite	1.73
CaO	0.74	0.82	Fe-Ti oxides	0.58
Na_2O	4.11	4.12		
K_2	4.69	4.70	Residual liquid	39.95
Total	99.69	99.74		

then:

$$C_L^* = \frac{C_0^*}{f} \text{ or } f = \frac{C_0^*}{C_L^*} \qquad (2)$$

Thus, as noted earlier the concentration of a hygromagmatophile element serves as an index of the degree of differentiation. Replacing equation (2) in equation (1) eliminates f and expresses the Rayleigh law in terms of the relative concentration of two elements:

$$C_L = C_0 \frac{(C_0^*)}{C_L^*} (\overline{D} - 1) \qquad (3)$$

This can be expressed in logarithmic form as:

$$\ln C_L = \ln C_0 + (1 - \overline{D}) \ln \left(\frac{C_L^*}{C_0^*} \right) \qquad (4)$$

This is a linear equation in $\ln C_L - \ln C_L^*$, if $\overline{D}$ remains constant throughout the process of fractionation. The slope of the line is given by $(1 - \overline{D})$. Theoretically, this allows the determination of $\overline{D}$ for any element. Examples of diagrams like these using samples from the Thomas Range are shown in Figure 4. A linear regression technique was used to determine the correlation and the slope of each variation line; the results are shown in Table 5. The elements Mg, Ca, K, Na, Sc, and Hf have correlation coefficients less than 0.85. Although such correlations may be statistically significant, they were not deemed useable.

The most critical factor that may affect the applicability of these derived $\overline{D}$'s is the assumption that $\overline{D}^U = 0$. Using the average modal mineralogy and mineral partition coefficients from Hildreth (1977) and Crecraft et al. (1981), the probable $\overline{D}^U$ in the lavas ranges from 0.06 to 0.16 (including up to 0.5% allanite or 0.04% zircon). Although these deviations may result in relative errors of about 0 to

20% in the range of f-values considered here (Allegre et al., 1977), they do not substantially affect the conclusions. In addition, the presumably low $\overline{D}^{Cs}$ and the correlation of Cs and U tend to support the assumption of small $\overline{D}$'s for both elements.

Using the $\overline{D}$'s of Table 5 and the element concentrations of Tables 1 through 3, f (the fraction of liquid remaining in a crystallizing system) can be calculated from equation (1). The results of these calculations, which use SM-61a as a parent, are shown in Table 6 and are compared with f's derived from the major-element mixing model. The calculated f's for each sample have a total range of about 0.1, except when using the data for F, La, and Th (in one instance). Although both the major and trace element models suggest that the rocks become more evolved in the sequence SM-61a, -62, -29-206, -42, there are large discrepancies between the f-values predicted by the two different types of calculations. Even including the possible 10 to 20% uncertainty produced by assuming the $\overline{D}$ for uranium equals 0, the two solutions appear exclusive. In fact to satisfy the f predicted by the major element model the $\overline{D}$ for U would have to be lower than 0 (an impossible situation), not higher as it may be in reality.

A series of equations may be defined from the derived $\overline{D}$'s and the observed phenocryst assemblage. These equations take the form

$$\overline{D}^i = D_{a/P}^i X_a + D_{b/P}^i X_b + \cdots$$

where $D_{a/P}^i$ = the partition coefficient of element i between phase a and liquid P, X_a = weight fraction of phase a. A set of these equations for various elements may be solved simultaneously to place limits on the composition of the fractionating mineral assemblage. We have attempted to do this using the mineral/liquid partition coefficients for high silica rhyolites of (Hildreth, 1977; Crecraft et al., 1981; Mahood and Hildreth, 1983). Solutions of these equations are generally consistent with the *relative proportions* of the major fractionating phases as

Table 5. Derived bulk distribution coefficients ($\overline{D}$) for rhyolites of the Thomas Range

	$\overline{D}$	r^2
Cs	-0.06	0.995
Be	0.11	0.986
Li	0.24	0.954
Lu	0.27	0.969
Ta	0.38	0.991
Rb	0.36	0.950
F	0.36	0.843
Th	0.50	0.998
Mn	0.63	0.868
Nb	0.70	0.859
Y	0.85	0.964
K	1.11	0.830
Zr	1.3	0.938
La	1.45	0.920
Ti	1.84	0.925
Sr	2.23	0.994
Eu	2.94	0.960

Note: Calculation of $\overline{D}$ explained in text; r^2 is a correlation coefficient obtained during linear regression of data.

Table 6. Comparison of f-values (residual liquid) derived from trace and major element models of differentiation

		f-value relative to SM-61		
Element	$\overline{D}$	SM-62	SM-29-206	SM-42
U	0	0.32	0.26	0.20
Cs	-0.06	0.34	0.26	0.18
Be	0.11	0.29	0.24	-
Li	0.24	0.40	0.24	-
Lu	0.27	0.48	0.20	0.22
Ta	0.38	0.34	0.25	0.20
Rb	0.36	0.33	0.27	0.18
F	0.36	1.42*	0.48*	0.16*
Th	0.50	0.17*	0.28	0.20
Mn	0.63	0.34	0.22	0.15
La	1.45	0.10*	0.04*	0.11*
Ti	1.84	0.44	0.19	0.26
Eu	2.94	0.41	0.21	0.21
Mean f**		0.37 ± 0.06	0.24 ± 0.03	0.20 ± 0.03
Major element model f		0.63	0.58	0.40

* Not included in computing mean.
** Mean ± 1 standard deviation.

determined from the major-element modeling (X sanidine ~0.46, plagioclase ~0.20, X quartz ~0.30, X biotite ~0.03, and X Fe–Ti oxides ~0.01). This mineral assemblage gives acceptable $\overline{D}$'s for Be, Cs, Rb, Sr, Mn, Li, Ti, Zr, Eu, and K. The La variation limits the maximum proportion of allanite (which effectively includes any monazite as well) to 0.0004 (X all = $\overline{D}/D_{allanite}^{La}$). The maximum proportion of fractionating zircon is limited to about 0.0004 by the Lu variation. The zirconium depletion of about 90 ppm implies a similar proportion (0.00043) of zircon (assuming f ~0.25 for the change). This amount would result in early cumulates with Zr concentrations of about 210 ppm and thus seems reasonable. The P depletion suggests that apatite made up about 0.00065 weight fraction of the cumulate (implying about 0.03% P_2O_5 in the cumulate).

Bulk distribution coefficients for Y, Nb, Ta, and Th calculated using this assemblage of minerals are significantly smaller than the $\overline{D}$'s derived from the trace element variation. This observation may simply imply that the individual mineral D's employed in the calculations are unrealistically low. It is not easy to explain $\overline{D}^{Ta}$ of $\overline{D}^{Th}$ in this fashion as they imply X zircon = 0.005 (or D_{Zircon}^{Ta} = 860) and X allanite = 0.001 (or $\overline{D}_{allanite}^{Th}$ = 1100). These values are 3 to 5 times larger than those implied by the REE variation noted above. Extremely small quantities of fractionating monazite (Mittlefehldt and Miller, 1983), titanite (identified by Turley and Nash, 1980, in some Thomas Range rhyolites), xenotime or even samarskite (reported from plutonic equivalents of topaz rhyolites in the Sheeprock Mountains of Utah, W. R. Griffitts, 1981, written communication) could account for the high $\overline{D}$'s calculated for Th, Y, Nb and Ta. Thus although we have not yet identified monazite or some of the other phases cited here, they have been reported from similar highly evolved magmas. The small quantities required would be extremely difficult to detect by petrographic examination. Alternatively, it is possible that some of the trace element variation is not produced by crystal/liquid fractionation—some liquid-state process may be operating. Thus the derived $\overline{D}$'s may be a measure of the "mobility" of different elements in the magma. Their variation could reflect their diffusion rates across gradients of P, T, and composition, their partitioning into an accumulating fluid, or their ability to form complexes with other migrating elements.

The lack of agreement between the f's derived from the major element and trace element models and the improbably high $\overline{D}$'s derived for some elements may indicate some sort of differentiation process in the liquid state. However, none of these observations suggest that crystal fractionation did not operate. Indeed the overall major and trace element systematics imply that any liquid differentiation was probably minor and may have only affected some trace elements. Keith (1982) and Nash and Crecraft (1981) came to similar conclusions regarding the differentiation of other rhyolitic rocks.

Liquid fractionation. The major processes of liquid differentiation that do not depend solely on crystal-liquid equilibria are (1) liquid immiscibility, (2) separation of a vapor or fluid, and (3) diffusion (thermogravitational diffusion, double-diffusive crystallization, the Soret effect).

The evolution of immiscible silicate and fluoride liquids is possible in fluorine-rich magmas (Koster van Groos and Wyllie, 1968; Kogarko and Ryabchikov, 1969; Hards and Freestone, 1978; Kovalenko, 1977b) and could theoretically play an important role in the trace element evolution of magmas. However, fluorine concentrations in hydrous alumino-silicate melts must be extremely high before immiscibility occurs (6 wt.% for granite–NaF–H_2O, Hards and Freestone, 1978; 3 to 4 wt.% for granite–HF–H_2O, Kovalenko, 1977b; 2.5 to 4 wt.% for ongonite–HF–H_2O, Kovalenko and Kovalenko, 1976). Such high levels of fluorine concentration are not reached in the Topaz Mountain Rhyolite (or the more fluorine-enriched rhyolites from the Spor Mountain Formation) and it appears that silicate liquid immiscibility played a negligible role in their evolution.

The separation of a fluorine-rich fluid and its consequent rise to the apical portion of a magma chamber could have had an effect on the evolution of topaz rhyolite magmas. It is difficult to assess the saturation concentrations of various F-species in igneous melts in the absence of direct experimental studies in H_2O-free systems. Based on the correlation of SiO_2 and F in ongonitic dike rocks, Kovalenko (1973) suggested that fluid-saturation occurs at 0.5 to 3 wt.% F, increasing with decreasing SiO_2 over the range of 75 to 69 wt.%. However, in view of the preference of F for melts over fluids (Hards, 1976) it seems more likely that H_2O would induce the saturation of a fluid which would contain some small amount of F. It thus seems appropriate to examine the H_2O–F–granite (or albite) system. Wyllie (1979) reports that F increases the solubility of H_2O in albite melts and Manning (1981) has suggested an imprecise lower limit for fluid saturation of 4 wt.% H_2O in H_2O–F–granite system. The late crystallization of biotite relative to quartz and feldspars and the experiments of Maaloe and Wyllie (1975) and Clemens and Wall (1981) in F-free systems suggest that H_2O concentration in these granitic melts may have been less than this. The position of the samples analyzed here in the Q–Ab–Or system is also consistent with the suggestion that the magma was not fluid-saturated. The emplacement of the rhyolites as lava flows instead of as pyroclastic deposits is also suggestive of a water-poor condition (Eichelberger and Westrich, 1981) and the absence of a free-fluid prior to venting. Hildreth (1981) has reviewed other evidence indicating that some magma chambers which give rise to large ash flows were not fluid-saturated. Nonetheless, Holloway (1976) has shown that even small quantities of CO_2 in granitic magmas will produce fluid-saturation. In the light of our ignorance of magmatic CO_2 concentrations it is worth pointing out that the

accumulation of a fluid near the roof of a magma chamber would result in net dilution of the magmatic concentrations of all trace elements except in the case of $D_{m/fl}$ (partition coefficient between melt and fluid) less than 1. Kovalenko (1977a) has shown experimentally that Li and Rb have fairly high $D_{m/fl}$ in an ongonite–H_2O–HF system at 1 kbar. With temperature decreasing from 800° to 600°C, $D_{m/fl}^{Li}$ decreases from 50 to 5 and $D_{m/fl}^{Rb}$ decreases from 250 to 50. Thus it seems impossible for $D_{m/fl}$ to attain values of less than 1 before a melt crystallized completely. Webster and Holloway (1980), Flynn and Burnham (1978) and Wendlandt and Harrison (1979) have shown that high $D_{m/fl}$ are common for REE in H_2O–Cl, H_2O–F, and CO_2-rich fluids at crustal pressures. On the other hand, Candella and Holland (1981) and Manning (1981b) have shown that Cu, Mo, and Sn prefer coexisting hydrous fluids ($\pm$F, Cl) over silicate melts. Thus a transient vapor-phase preceding or coincident with eruption may have altered the concentrations of these elements but probably did little to effect the concentrations of other elements including the REE.

If element enrichment/depletion patterns can be considered indicative (and there is no evidence that it is so simple), the apparent similarity of the variations in the rhyolites from the Thomas Range to those of the Bishop Tuff (Hildreth, 1979) could be taken as evidence that "convection-aided thermogravitational diffusion" produced some of the characteristics of the topaz rhyolites (Fig. 5). Samples SM-41 or SM-29-206 could represent eruptions from the upper part of a magma chamber zoned at depth to a composition similar to SM-61a. However, the uncertainty in the temporal relationship among the sampled lava flows as well as the rapid reestablishment of chemical zonation in some magma chambers following their eruption (Smith, 1979), precludes postulating that the lavas were erupted from a unitary system like the Bishop Tuff magma chamber. The samples may simply record temporal variations in a magma chamber like those described by Mahood (1981) and Smith (1979). Whether the process which gave rise to these features (Fig. 5) was thermogravitational diffusion (Shaw et al., 1976; Hildreth, 1979) or crystallization in a double-diffusive system with the rise of evolved less-dense liquid from the walls of a crystallizing magma chamber (Chen and Turner, 1980; McBirney, 1980; Nash and Crecraft, 1981; Christiansen, 1983), or some other process is difficult to evaluate until more quantitative formulations of these hypotheses can be made.

In any case, fluorine probably played an important role in the differentiation process because of its tendency to form stable complexes with a number of elements, including the REE (Smirnov, 1973; Mineyev et al., 1966; Bandurkin, 1961). Halogen complexing (and subsequent transport in a vapor-phase, or within a melt, down thermal, chemical, or density gradients) has been invoked to explain the incompatible trace-element enrichment of evolved magmas and pegmatites (Bailey and Macdonald,

1975; Fryer and Edgar, 1977; Taylor et al., 1981; contrast with Muecke and Clarke, 1981) and the depleted nature of some residual granulites (Collerson and Fryer, 1978). For the REE, the stability of the fluoride complexes increases from La to Lu and at lower temperatures (Moller et al., 1980). Mineyev et al. (1966) experimentally demonstrated the increased mobility of Y (as representative of HREE) relative to LREE (La and Ce) in acidic solutions with high fluorine and sodium activities across a temperature gradient. Thus, the pronounced HREE enrichment (considered to be a fingerprint of liquid fractionation in rhyolitic magmas) during the evolution of the Topaz Mountain Rhyolite may have been the result of the co-migration of F with HREE (and presumably with Li, Be, U, Rb, Sn, etc.) to the apical portions of a magma chamber. From studies of rhyolitic ash-flow tuffs, Mahood (1981) has suggested that the dominant control on ash-flow trace-element enrichment patterns may be their roofward migration as volatile complexes. The most important volatiles in highly evolved granitic magmas are probably H_2O, HF, HCl, B, and CO_2(?). The suite of elements associated with these volatiles should differ depending on the nature of the complexes or ion-pairs formed by each volatile (Hildreth, 1981).

In another paper (Christiansen et al., 1983b), we suggest that the trace element evolution of rhyolitic magmas is sensitive to F/Cl in the melt. With differentiation, F-dominated systems (F/Cl greater than 3) display increases in Al, Na, Li, Rb, Cs, Ta, Th, and Be, while Cl-dominated systems are usually peralkaline and are associated with greater enrichments of LREE, Na, Fe, Ti, Mn, Zn, Nb, and Zr. Although magmas with variable proportions of H_2O, F, Cl and CO_2 certainly exist, topaz rhyolites appear to represent the fluorine-dominated end member. However, it is important to note that the volatiles and alkalies also affect the stability of various mineral phases important to the chemical evolution of these rhyolites (allanite, zircon, biotite, titanite, apatite, pyroxene, and hornblende). Thus, it is not an easy task to distinguish liquid-state processes from those that result from fractional crystallization of a distinctive mineral assemblage which was stabilized by different volatile-element concentrations or ratios.

We suggest that the extreme compositions observed in lavas from the Topaz Mountain Rhyolite are the result of both extensive fractional crystallization and at least to some extent the result of the diffusive transfer of trace elements. Although some of the magmas may have been fluid-saturated shortly before eruption, the role of a separate fluid phase in their chemical evolution appears to have been minor.

Spor Mountain Formation

The limited number of trace element analyses for samples of the Spor Mountain Rhyolite preclude detailed petrochemical modeling of its differentiation. However, comparisons with the younger Topaz Mountain Rhyolite

and with Asian "ongonites" suggest some ideas about its evolution.

The composition of the Spor Mountain Rhyolites implies that they evolved from magmas slightly different from those erupted in the Thomas Range. They are richer in Fe, Cr, Co, Ba, LREE, and depleted in Si relative to the evolved Thomas Range lavas, trends which are inconsistent with the continued differentiation of the Topaz Mountain Rhyolite. In contrast, they are extremely enriched in U, Rb, Th, Cs, Li, Be, Sn, Ta, and the HREE. The trace element compositions do not lie on the trends defined by the Topaz Mountain Rhyolite, implying that $\overline{D}$'s were different, that initial concentrations varied or both. However, the overall chemical and mineralogic similarity of the two sequences (REE patterns, enrichment and depletion of the same elements relative to other rhyolite types, Fe-enriched mafic minerals) suggests that the Spor Mountain magma and the rhyolites of the Thomas Range may have evolved from grossly similar parents. The most important features of the older rhyolites that require explanation are (1) the low Si and high Al contents, (2) Na/K greater than 1, and (3) the extreme enrichment of "fluorophile" elements.

The high Na/K (relative to other topaz rhyolites) of the Spor Mountain Rhyolites can be explained in terms of the effect of fluorine on residual melt compositions in the granite system (Manning 1981; Kovalenko 1977b). Both sets of experiments show a regular increase in the size of the quartz field in the Q–Ab–Or system and increasingly albitic residual melts with increased fluorine content (Fig. 2). Thus as the activity of the fluorine increases in a residual melt, either by differentiation or by the isochoric crystallization of a fluid-saturated melt (Kovalenko, 1973), quartz fractionation becomes more important leading to a reversal in the normal trend of SiO₂ increase with chemical evolution. As a result aluminum increases in the residual melt and continued fractionation of sanidine leads to higher Na/K in the liquid. This move to more sodic liquid compositions as a result of potassium feldspar fractionation is petrographically manifest by the common mantling of plagioclase by sanidine (anti-rapikivi) in the lavas from Spor Mountain.

Each of these trends (decreasing SiO₂ and increasing Na and Al with higher F) is observed in the extremely F-rich topaz rhyolites like those at Honeycomb Hills, Utah (Christiansen et al., 1980) and are well developed in F-rich ongonites from central Asia (Kovalenko and Kovalenko, 1976). These trends are also mimicked during the crystallization of rare-metal Li–F granites, strengthening the suggestion that topaz rhyolites may be associated with mineralized plutons at depth.

The Spor Mountain Rhyolite appears to represent the product of protracted fractional crystallization under the influence of high fluorine activity in the melt. Liquid fractionation is suggested, but not demonstrated, by the flat REE patterns (Fig. 3). Although the magma may not have been fluid-saturated during the most of its evolution,

the association of numerous tuff-lined breccia pipes (Lindsey, 1982) with the Spor Mountain magmatism suggests that the magma became saturated during eruption as a result of a rapid pressure drop. Enhanced vapor-phase transport of Sn and Mo and perhaps Be, Li, Cs, U, and other elements may have occurred under these conditions adding to the effect of other differentiation processes. This mechanism could lead to the rapid formation of a lithophile-element rich cap to the magma chamber. The subsequent eruption of the volatile-rich cap may be represented by the emplacement of the Be (Li–F–U)-mineralized tuff that underlies the Spor Mountain rhyolite.

Summary

The fluorine-rich rhyolites of the Thomas Range and Spor Mountain, Utah, are enriched in a distinctive group of lithophile elements (U, Th, Rb, Li, Be, Cs, Ta and Sn) typical of other topaz rhyolites from the western United States. The mineralogy and geochemistry of the lavas suggests that they are similar to "ongonites", aluminous bimodal rhyolites, and anorogenic granites.

The topaz rhyolites from the Thomas Range are interpreted as being derived by differentiation from a less silicic but still rhyolitic magma. Calculations suggest that sanidine, quartz, sodic plagioclase, Fe-Ti oxides, biotite, allanite, zircon and apatite were the dominant fractionating phases over the course of differentiation sampled. However, failure of fractional crystallization models to explain the high bulk partition coefficients of Ta and Th suggests that liquid fractionation, aided by the high fluorine content of the magmas may have occurred *or* that some unidentified phase (monazite, titanite, samarskite, xenotime) was removed from the evolving melt.

The Spor Mountain rhyolite contains higher concentrations of fluorine and lithophile elements than the Topaz Mountain Rhyolite. It may have evolved from a magma generally similar to the Topaz Mountain Rhyolite, but its differentiation was more protracted. Expansion of the quartz stability field by elevated fluorine concentrations and the resultant separation of quartz and sanidine may have led to its lower Si and higher Na and Al contents. Extended fractional crystallization, possibly aided by liquid fractionation, led to extreme concentrations of Be, Li, U, Rb, Cs and other fluorophile elements. The magmatic concentrations of these elements set the stage for the development of the Be-mineralized tuff which underlies the lava.

Acknowledgments

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