Volcanology and Petrology of the Rattlesnake Ash-Flow Tuff, Eastern Oregon

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Title: Volcanology and Petrology of the Rattlesnake Ash-Flow Tuff, Eastern Oregon.

Abstract approved: ___________ Anita L. Grunder

The Rattlesnake Ash-Flow Tuff erupted 7.05±0.01 Ma from the western Harney Basin, southeastern Oregon. The location of the vent area is inferred based on vent-ward increases in size of pumices, in degree of welding, and in degree of postemplacement crystallization. Today's outcrops cover 9000 km² and estimated original outcrop coverage was ca. 35000 km². Tuff thickness is uniform ranging mainly between 5 and 30 m with maxima up to 70 m. Estimated eruption volume is 280 km³ dense rock equivalent (DRE). Lithological variations include vitric non- to densely welded tuff, vapor-phase, devitrified, spherulitic, lithophysal, and rheomorphic tuff. Lithological zoning characteristics of the tuff change locally at nearly constant tuff thickness over distances of 1 to 3 km grading from incipiently welded tuff to highly zoned sections. Regional variations become apparent by integrating many sections from one area. A three-dimensional facies model is developed describing the local and regional facies variations.

The Rattlesnake Tuff consists of high-silica rhyolite (HSR) erupted as pumices and glass shards. Dacite pumices make up less than 1% of the total volume and quenched basalt and basaltic andesite inclusions inside dacite pumices constitute << 0.1 volume %.

HSR pumices cluster in 4 to 5 compositional groups which are discerned best by La, Eu, Ba, Hf(Zr), Ta(Nb). Major element variations are minor but consistent between groups with SiO2 increasing and FeO*, MgO, TiO2, and CaO decreasing with differentiation. Modal mineralogy, mineral chemistry, and partition coefficients also change progressively. The diversity of HSR is likely the product of crystal fractionation processes. A model is proposed by which a stratified magma chamber is generated from the roof of the chamber downward by progressively more evolved HSR. Least evolved HSR is likely the product of dehydration melting of high-grade intermediate to mafic protoliths. High Ba/Rb ratios of 30 in the least evolved HSR limit the potential protoliths to lithologies with equal or higher Ba/Rb.

Dacite pumices and mafic inclusions reveal the nature of the mafic root zone to the Rattlesnake Tuff magma chamber. Quenching and mingling textures with the host pumice indicate that basaltic andesite and basalt inclusions are cognate. Dacite formed at the interface between HSR and underlying enriched basaltic andesite and was generated by mixing of these two components. Basaltic andesite is fractionated and enriched in trace elements compared to regional primitive tholeiite (HAOT). Basaltic andesite has evolved from HAOT's mainly through fractionation and recharge while being stalled underneath a silicic cap. APPROVED:

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Dean of Graduate School

Date thesis is presented _____ April 27, 1994

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Typed by the author <u>Martin J. Streck</u>

An honest tale speeds best being plainly told.

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King Richard III ACT IV

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Preface

The dissertation consists of three chapters of which each is designed as a separate publication. All chapters concern various aspects of the Rattlesnake Ash-Flow Tuff erupted 7 Ma ago from the Harney Basin in southeastern Oregon. The author of the thesis is the first author on the first two papers (Chapter I and II) and is the only author of the last paper (Chapter III). Data acquisition, compilation, and interpretation were done mainly by the thesis author.

Chapter one addresses volcanological aspects. First, data are presented concerning the distribution and thicknesses of the Rattlesnake Tuff and is followed by estimates of original areal coverage, the volume of magma erupted, and the likely vent area. The main part of the first chapter is concerned with the appearance and distribution of welding and crystallization facies in the Rattlesnake Tuff and with development of a facies model.

The second and third chapters are concerned with petrologic aspects of the Rattlesnake Tuff, which consists of mainly high-silica rhyolites, but also includes dacites and basaltic components. In particular, the second chapter presents petrographic, mineralogical, and chemical data concerning the chemical and physical relationships of basaltic and dacite compositions to the rhyolite and to each other.

The third chapter treats in detail the diversity of high-silica rhyolites as recorded by compositional pumice clusters. Chemical, mineralogical, and mineral-melt partition coefficient data are presented. This chapter discusses the likeliest differentiation process responsible for progressively more evolved high-silica rhyolite magmas.

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Volcanology and Petrology of the Rattlesnake Ash-Flow Tuff, Eastern Oregon

Chapter I: Facies variations in a thin, widespread ignimbrite sheet: The Rattlesnake Tuff, Eastern Oregon.

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Abstract

Today's outcrops of the Rattlesnake Ash Flow Tuff, emplaced 7.05 my ago, cover ca. 9000 km², but reconstructed original coverage was between 30,000 and 40,000 km². Thicknesses are remarkably uniform, ranging between 15 to 30 m for the most complete sections. Only 13% of the outcrop area is covered with tuff thicker than 30 m, up to a maximum of 70 m. Present-day estimated tuff volume is 130 km³ and reconstructed erupted magma volume is 280 km³ dense rock equivalent (DRE). An exponential decrease in average pumice size implies a source area in the western Harney Basin, centered on the main outcrop areas. Distances from inferred source to most distal outcrops are 150 km. A large number of welding and crystallization zones formed during post-emplacement processes. The zone of partial welding, the vapor phase zone, and the lithophysal zone are divided into macroscopically distinguishable subzones. Rheomorphic vitric to devitrified tuff is found up to a radius of 40-60 km around the inferred source. In non-rheomorphic tuff, lithological zoning of individual sections, at constant outcrop thickness of 17-23 m, varies from vitric, non- to incipiently welded tuff throughout to highly zoned sections consisting of a basal non- to densely welded vitric tuff overlain by a zone of crystallized tuff which grades internally from spherulite to lithophysae dominated, to a zone of devitrification, and finally to a zone of vaporphase crystallization and is capped by upper partially welded vitric tuff. The two facies zonation extremes occur within 1-3 km of one another at several places. Welding and

crystallization decrease with distance from the inferred source, but the regional pattern becomes apparent only by integrating many sections within a given area. Strong local variations are interpreted to be the result of threshold-governed welding and crystallization near the critical welding temperature due to slight original thickness differences influencing the thermal insulation. A three-dimensional welding and crystallization facies model has been developed based on 85 measured sections incorporating local and regional variations.

Introduction

Zones and zonal variations in ignimbrites are based mainly on the classic works of Smith, (1960a, b) and Ross and Smith (1960). Description of horizontal variations for single extensive ignimbrite sheets were rare 30 years ago (Smith, 1960a, p.149). Despite the long time span since, relatively few reports on horizontal variations of welding and crystallization zones in welded tuffs have been added. This study focuses on the lateral and vertical variations of welding and crystallization zones in the Rattlesnake Ash-Flow Tuff, an important time and structural marker in eastern Oregon, USA (Walker, 1979). The tuff is extremely widespread, almost evenly thin, and well exposed. Welding, crystallization, and rheomorphic processes in the Rattlesnake Tuff are responsible for a large number of facies with excellent preservation. The tuff includes features characteristic of non-welded to lava-like ignimbrites (e.g. Branney & Kokelaar, 1992). A facies model based on the Rattlesnake Tuff documents the complexities that one might encounter in other comparable welded, thin, and widespread ignimbrites, such as are typical of the Oregon Plateau and adjacent areas in Idaho (Mansfield & Ross, 1935, Stearns & Isotoff, 1959, Walker, 1970 & 1979, Greene, 1973, Bonnichsen & Citron, 1982, Ekren et al., 1984). The broader importance of zones and zonal variations in ignimbrites is the constraint they place on eruption and emplacement dynamics of large-scale volcanic processes that have never been witnessed.

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Tectonic setting

The late Miocene Rattlesnake Ash-Flow Tuff (hereafter referred to as the Rattlesnake Tuff) is exposed over at least 9000 km² in eastern Oregon, which can be broadly divided into two main tectonic regions. The Blue Mountains Province, to the north (Fig. I-1), is characterized by almost continuous volcanism interrupted only by short periods of deformation and uplift since the Eocene (Fisher, 1967, Swanson & Robinson, 1968, Robyn & Hoover, 1982, Robinson et al., 1984, Hooper & Swanson, 1990, Walker & Robinson, 1990). Compressional features prevail, although normal faulting is also part of the structural record (Robyn and Hoover, 1982, Barrash et al., 1983, Robinson et al., 1990, Walker & Robinson, 1990). The northern Basin and Range Province, to the south, is characterized mainly by middle Miocene to Quaternary volcanic lithologies that are cut by high-angle, NNE-striking normal faults bounding a few large fault blocks and zones of NW-trending faults (Lawrence, 1976, Walker & Nolf, 1981, McKee et al., 1983). Volcanic rocks as old as Middle Tertiary are exposed along the Steens and Hart Mountain fault blocks (Walker, 1979). The NW-trending Brothers Fault Zone is the transition between the northern Basin and Range and the Blue Mountains Province and broadly coincides with the High Lava Plains physiographic province (Fig. I-1) (Lawrence, 1976).

The High Lava Plains are mainly Late Miocene - Pliocene volcanic rocks consisting dominantly of high-alumina olivine tholeiite lava flows; silicic volcanism produced a few thin, but widespread ignimbrites and numerous silicic dome complexes (Walker, 1970, Walker, 1974, MacLeod et al. 1976, Hart et al., 1984, Carlson & Hart, 1987, Draper, 1991). Silicic magmatism progressed in age from around 10 Ma near the southeastern end of the Brothers Fault Zone to less than 1 Ma near the western termination, based on radiometric ages of silicic domes (Fig. I-1) (MacLeod et al., 1976).

Fig. I-1 Regional overview

Regional setting and outcrop pattern of the Rattlesnake Tuff. Light stipple represents Blue Mountains Province, dense stipple shows Rattlesnake Tuff; proposed source area indicated by solid circle. Dotted lines with numbers are simplified isochrons in million of years for NW-migrating silicic volcanism, after MacLeod et al. (1976). Solid lines indicate faults and stars indicate Cascade stratocones.



	Age d	lata for sing	le feldspar cr	ystals	
Lab ID#	Ca/K	36 _{Ar/} 39 _{Ar}	40 _{Ar*/} 39 _{Ar}	%40Ar*	Age(Ma) ± 1σ
Sample: pur	nice HP91-	12			
5357-01	0.302	0.00014	0.778	93.8	6.99±0.0
5356-06	0.096	0.00013	0.783	93.2	7.03±0.0
5357-03	0.098	0.00031	0.783	87.8	7.03±0.0
5357-04	0.082	0.00010	0.784	94.2	7.04±0.0
5357-07	0.110	0.00012	0.784	93.5	7.04±0.0
5356-07	0.085	0.00014	0.784	92.9	7.04±0.0
5356-08	0.063	0.00011	0.784	93.9	7.04±0.0
5356-01	0.062	0.00010	0.785	94.0	7.04±0.0
5357-02	0.105	0.00007	0.785	95.2	7.04±0.0
5356-04	0.082	0.00011	0.786	93.9	7.05±0.0
5356-05	0.111	0.00031	0.786	87.8	7.05±0.0
5356-03	0.025	0.00005	0.786	95.6	7.06±0.0
5357-05	0.099	0.00013	0.787	93.4	7.06±0.0
5357-06	0.076	0.00010	0.787	94.2	7.07±0.0
5356-02	0.036	0.00008	0.787	94.5	7.07±0.0
				Wtd.Ava.	7.05±0.0

Table I-1 40Ar-39Ar analytical data for the Rattlesnake Tuff

Notes: Errors in age quoted for individual runs are 1 σ analytical uncertainty. Weighted averages are calculated using the inverse variance as the weighting factor (Taylor, 1982), while errors in the weighted averages are 1 σ standard error of the mean and incorporate error in J (see below) (Samson & Alexander, 1987). Ca/K is calculated from 37 Ar/ 39 Ar using a multiplier of 1.96. 40 Ar* refers to radiogenic argon.

 $\lambda = 5.543 \times 10^{-10} \text{ y}^{-1}$.

Isotopic interference corrections: $({}^{36}\text{Ar}/{}^{37}\text{Ar})_{Ca} = 2.58 \times 10^{-4} \pm 6 \times 10^{-6}$, $({}^{39}\text{Ar}/{}^{37}\text{Ar})_{Ca} = 6.7 \times 10^{-4} \pm 3 \times 10^{-5}$. $({}^{40}\text{Ar}/{}^{39}\text{Ar})_{K} = 2.19 \times 10^{-2} \pm 2 \times 10^{-4}$.

 $J = 0.004986 \pm 0.00001$. Age determinations were done by Alan Deino at the Geochronology Center of the Institute of Human Origins, Berkeley, California.

Name and age

The Rattlesnake Tuff crops out prevalently in a few areas separated by regions where it is absent, or exposed only in patches. Early workers divided this vast ignimbrite sheet into different units based on localities, namely, the Double O Ranch Tuff, Twelvemile Tuff, and Rattlesnake Tuff (Green et al. 1972, Enlows, 1972, Thayer, 1952). On stratigraphic, petrographic and chemical grounds, Beeson (1969) and Davenport (1971) argued for a single tuff unit.

The Rattlesnake Tuff has been radiometrically dated several times by potassiumargon methods, yielding an average age of 6.5 Ma (see Walker, 1979, for summary). Recent high precision 40Ar/39Ar radiometric age determinations on fifteen single crystals separated from one pumice yielded a weighted average of $7.05 \pm 0.01(1 \text{ sigma})$ (Table I-1), which overlaps the oldest reported K-Ar ages. The radiometric age determinations were performed by Alan Deino at the Geochronology Center of the Institute of Human Origins, Berkeley California.

General aspect

The Rattlesnake Tuff consists of a single cooling unit throughout the outcrop area, where it typically occurs as 10 to 20 m thick plateau-capping rimrock. Evidence for as many as 3 flow units was found only at one location (100 km NNW of Burns, near Dayville, Fig. I-1). There, a clear separation into different flow units is apparent through sharp interfaces between nonwelded zones each 1-2 m thick. The top of the section is made up of surge deposits (see Fig. I-2) which originated very likely through secondary ash-flows (M. Rosi, personal comm., 1993). Other evidence for two flow units elsewhere is based on reversal in the proportion of various glass shard populations. A sharp interface between upper crystallized tuff and lower vitric tuff was mistakenly interpreted by Parker (1974) as separate flow units south and west of Harney Lake.

The Rattlesnake Tuff is nonwelded to densely welded with superimposed vapor-phase, pervasive, spherulitic, and lithophysal crystallization zones. Phenocryst content is around 1% or less for the bulk tuff. The tuff is pumiceous with distinctive varieties of pumice clasts. Pumices are white, gray, black or have banded textures. A range of high-silica rhyolite (75-77.5 wt.% SiO₂) and dacite compositions (62-70 wt.%

SiO₂) is recorded in the different colored pumices. Glass shards of the matrix of the tuff can also be separated into different populations based on color. The glass shards include only high-silica rhyolite compositions. The number of pumices ranges from $\pm 40\%$ near source to 2% distally (Fig. I-3).

Distribution, thickness and volume

Today's outcrops

Occurrences of Rattlesnake Tuff were compiled from a variety of maps, publications, and theses and were checked and supplemented through field work (Fig. I-4). The main outcrop areas of the Rattlesnake Tuff are southern and northern Harney Basin, Paulina Basin, and John Day Valley. Trace element data, determined by instrumental neutron activation analysis (INAA), for samples from the most distal outcrops support the previously established correlation for these areas (Fig. I-4) (Johnson, 1960; Wilcox & Fisher, 1966; Smith et al., 1984; Enlows, unpub. data).

Thickness information from 240 sections was compiled into a present-day isopach map for the whole tuff sheet (Fig. I-5). The Rattlesnake Tuff is semicontinuously exposed over an area of ca. 9250 km² (Table I-2, Fig. I-4, I-5). The area covered by tuff thicker than 30 m is 1244 km² and describes a southern and a northern arc distributed around the inferred source area (discussed below). The rest (87%) of the total area is covered by tuff with a thickness of less than 30 m. Most of the tuff is remarkably uniform in thickness, between 5 and 30 m, despite its wide extent. Individual sections as thick as 30 m were mainly established by cliff measurements with rope; thicker sections were determined either by hand-leveling or were estimated from contours on a 7.5 minute topographic map. Most thickness determinations are minima mainly because of erosion effects. Based on lithology and crystallization facies, however, at most a few meters are missing in most cases. Throughout the distribution area, the lower boundary of the tuff crops out at numerous localities (Fig. I-3c, I-8). When the base was not exposed, it could be constrained by comparison of the thicknesses of zonal variations in similar, but more complete sections. Constraints on the original upper boundary are more ambiguous. Upper nonwelded tuff is preserved in only one locality. However, highly zoned sections (Fig. I-3c) commonly include upper partially welded, vitric tuff, suggesting proximity on the

Fig. I-2 Surge deposits

Local development of waveforms (originated likely through secondary ash flows) in John Day Valley, 114 km from the inferred vent; flow direction was from left to right; note, deposition of basalt lithic fragment lens on lee side of chute and pool structure (upper right).



Fig. I-3 Lithologic variations

a) (Top): Pumice-rich tuff with 30 to 50% pumices, 37 km from inferred vent. Outcrop yielded an average of five largest pumices of 57.4 cm. Plastic ruler is 12 cm in length.
b) (Bottom): Pumice-poor (ash-rich) section, 148 km from inferred vent, containing ca. 2% pumice lapilli. Maximum average lapilli size for this locality is 3.7 cm. Coin for scale is 1.9 cm.



Fig. I-3, Continued:

c) Section of Rattlesnake tuff, 16 m thick with thin, laminated basal ash-fall (white). From base to first ledge, vitric tuff grades from nonwelded to densely welded; section above first ledge to ca. 3 m below top consists of lithophysal zone. Top 3 m are of pervasively devitrified zone; notice that hackly jointed underlies massive, ledge-forming devitrified tuff.



Fig. I-4 Outcrop map

Outcrop map with reconstructed original extent. See text for explanation of reconstruction categories 'certain' to 'probable'. Insert shows area of figure I-4 relative to Oregon. Arrows point to distal tuff localities correlated with Rattlesnake Tuff based on trace-element concentrations (see text). Outcrop pattern in John Day Valley shows Rattlesnake Tuff and closely related conglomeratic units of Brown & Thayer (1966). Sources are: Brown & Thayer (1966), Davenport (1971), Greene et al. (1972), Johnson (1960), Swanson (1969), Walker (1963), Walker et al. (1967), Walker & Repenning (1965), Wallace & Calkins, J. A. (1956), Wilcox & Fisher (1966).

g. I-4:



Fig. I-5 Isopach map

Thickness of the Rattlesnake Tuff is shown in meters. Dashed circle is inferred source area. Most distal tuff remnants not shown. Insert shows area of figure I-5 relative to Oregon.





order of 1 to 3 m from the top of the noneroded tuff. This estimate assumes that the original upper non- to partially welded, vitric tuff was comparable in thickness to the basal vitric zone (c.f. Riehle, 1973). Even if upper vitric sections were twice as thick as basal ones, the thickness estimates would not be strongly affected. Exposure thicknesses of less than 10 m in strongly welded or crystallized tuff (see regional facies variations) within a radius of 80 km from the inferred source are very likely due to erosion and do not represent nearly complete sections.

Original areal coverage

Reconstruction of the area originally covered with tuff has important implications for estimates of the volume erupted. The estimate of original area coverage consists of the area covered by today's outcrops (\approx area of isopach map) and the reconstructed area originally covered by tuff but later eroded or buried by younger deposits. The reconstructed area was divided into four qualitative categories, "certain", "very likely", "likely" and "probable", indicating increasing degrees of uncertainty in the estimate (Fig. I-4, Table I-2). "Certain" includes very proximal areas around the vent and places well surrounded by today's outcrops. "Very likely" are proximal areas or areas fairly well surrounded by present outcrops. "Likely" was assigned to lobes needed to connect well correlated, but very distal remnants of the tuff or areas not well surrounded by outcrops. The solid line in figure I-4 encloses the cumulative area of the categories "certain", "very likely", and "likely". The category "probable" includes areas that make for a more equant distribution of the Rattlesnake Tuff and the limit is indicated by the dashed lines in figure I-4. The estimate of the original area covered with Rattlesnake Tuff is 35,000 km², which might be still a minimum value (Table I-2). The above estimate of the original coverage is based on two major inferences: (1) the Rattlesnake Tuff represents the deposit from high-energy ash-flows; and (2) the relief throughout most of the deposition area was low.

The energy of an ash flow is reflected by the geometry of its deposit (Walker, 1983). The aspect ratio (average tuff thickness / diameter of the circle enclosing all outcrops) of the Rattlesnake Tuff can be used to infer the high-energy character. Aspect ratios calculated for the Rattlesnake tuff range from 2.5×10^{-4} to 5×10^{-5} using very conservative to optimistic values of 50 m/200 km and 15 m/300 km, respectively. Both aspect ratios fall within the range of areally extensive high-energy ash flow tuffs (c.f. Fig. 1 in Walker, 1983).

Low relief at the time of eruption of the Rattlesnake Tuff is indicated by the uniform thickness of the tuff which is between 15 and 30 m for nearly complete sections throughout most of the outcrop area and its landscape-mantling appearance where it is continuously exposed. A more rugged topography would have generated thicker valley deposits thinning outward with a more patchy distribution pattern as described from the Campanian ignimbrite (Fisher et al., 1993). Paleotopography induced thickening of the tuff deposit is possibly indicated by the distribution of the deposit thicker than 40 m. The two areas (Fig. I-5) are located on the projected continuation of the lee side of older volcanic edifices (Dry Mountain, Wilson Butte). Other likely topographic barriers causing either very thin or no deposit were the Strawberry and Aldrich Mountains (see Fig. I-4) and local areas around older domes or composite volcanoes.

Volume

Previous volume estimates of the Rattlesnake Tuff spanned a wide range from 10 km³ (Draper, 1991), to 150-200 km³ (Walker, 1970), to 1500 km³ (Parker, 1974). Using the established isopach map with conservative average thicknesses, the volume preserved in the present day outcrops is 132 km³ (Table I-2). The tuff volume for the reconstructed area was calculated using conservative thicknesses comparable to nearby sections. The total volume (present-day + reconstructed) is 332 km³ (Table I-2). Because much of the tuff is not densely welded, the magmatic volume is less. Dense glassy rock from the Rattlesnake Tuff has a specific gravity of 2.34 g/cm³. A range of specific gravities, based on measurements of representative samples, was used for the different isopach areas to approximately account for welding variations. The total magma volume is 281 km³ (Table I-2). No air-fall tuff or possible intra-caldera ignimbrite is considered, making this a minimum estimate.

Table I-2 Area and volume estimates

	A	rea and Vo	lume Est	imates of	Rattlesna	ke Tuff		
	Area	[km ²]			Volume [km ³]			
				Т	uff	pt/pg1	Magma	erupted¥
isopach area	a:	cumulative	Th¶		cumulative			cumulative
5-10m	4845	4845	5	24.2	24.2	0.77	18.6	18.6
10-20m	1535	6380	15	23.0	47.2	0.86	19.8	38.4
20-30m	1624	8004	25	40.6	87.8	0.94	18.2	56.6
30-40m	591	8595	30	17.7	105.5	1	17.7	74.3
> 40m	653	9248	40	26.1	131.6	1	26.1	100.4
sub-total	9248			131.6			100.4	
reconstruct	ed:							
certain	3322	3322	10-15	46.3	46.3	0.9	41.7	41.7
very	5042	8364	8-12	52.1	98.4	U	46.9	88.6
likely								
likely	7683	16047	5-10	51.1	149.5	11	46	134.6
probable	9701	25748	5-8	50.8	200.3	н	45.7	180.3
total	34996			331.9			280.7	

I conservative thicknesses [m] used for volume calculations

 ρ^{t}/ρ^{gl} = density of tuff/ density of dense rock (ρ =2.34) [¥] erupted magma calculated as dense rock equivalent with ρ^{t}/ρ^{gl} ratio

Fig. I-6 Average maximum pumice and lithic fragment

(a) Average maximum pumice. Values are in [cm] calculated from the five largest pumice clasts and are contoured approximately. Dashed circle is the inferred source.(b) Average maximum lithic fragment, in cm. Refer to figure I-7b for the number of averaged lithic fragments for each site. Framed numbers indicate sites with a lithic enriched zone near the base of the tuff, see text for details.

Fig. I-6:



Fig. I-7 Distance versus maximum pumice and lithic fragment

(a) Distance versus average maximum pumice. \mathbb{R}^{2} 's compare exponential to a linear regression fit. Solid line indicates exponential fit through data points, values with half-filled squares, representing poor outcrops, were omitted for evaluation in figure I-9. (b) Distance versus average maximum lithic fragments. Circles represent sites with averages calculated from 4 or 5 lithic fragments, squares are sites where ≤ 3 lithic fragments were found and averaged; circles with crosses are sites with basal lithic enriched zone. Solid line represents linear regression fit to the data set of the circles.



Depositional facies

Pumice clasts

The Rattlesnake Tuff is pumiceous in outcrops where strong welding and(or) crystallization did not obliterate the original pumiceous texture. Pumice clast investigations were undertaken at outcrops only incipiently welded throughout or in thicker non- to partially welded sections where pumices are not deformed by welding (Fig. I-3a & b). The Rattlesnake Tuff contains different pumice clast populations which can be divided into homogeneously colored white, gray, black pumices and variously colored banded pumices of any combination of the white, gray and black type; unbanded black pumices are exceedingly rare. White and gray pumices are metaluminous high-silica rhyolite; the black pumices include high- to low-silica, slightly alkalic dacite compositions.

The largest pumices at a given outcrop are always of the white and gray type. There are outcrops, usually beyond 80 km from the inferred source, which consist almost exclusively of white pumices; others consist of about equal proportions of white, gray and mingled pumices. Typically, but not exclusively, the abundance of gray and black pumices increases upward within the first one or two meters. In general, the proportion of white pumices is greatest at the base and at distal outcrops. The vertical and lateral variations in different colored shard populations mimic the variations in pumice populations. These variations can be interpreted as tapping of successive deeper levels of a zoned magma chamber (Smith, 1979) as a first order process, superimposed on by deposition and erosion mechanics of overlapping ash flows.

The results from calculating the average sizes of the 5 largest pumice clasts at many localities show a decrease of average maximum size away from the inferred source (Fig. I-6a and I-7a). Pumice clast averages range from 57.4 to 1-3 cm. The rate of decrease follows an exponential curve rather than a linear one (Fig. I-7a). No vertical grading of pumices was observed, except for the lack of largest pumices within 1 m of the base of the tuff. A decrease of average pumice size with distance has been found in some pyroclastic flow deposits (Yokoyama, 1975, Kuno et al., 1964, Wright & Walker, 1977, Fisher, 1966). However, in many tuffs there is no simple relationship of average pumice size with distance; in some cases, size and distance correlate positively (Cas & Wright, 1987, p.194).

Along with the decrease of average maximum pumice size with distance from the source, there is a decrease of the abundance of pumices. This is exemplified by comparing figure I-3a & b, which represent proximal and distal outcrops, respectively. The proportion of pumices ranges from ca. 30% -50% in proximal localities to 2% distally (Fig. I-3a & b). Thus, the Rattlesnake Tuff changes its character laterally from being a pumice-rich tuff to an ash-rich tuff.

Lithic fragments

Accidental lithic fragments make up one percent or less of the tuff. Lithic fragment assemblages are dominated by basalt to basaltic andesite and are sometimes monolithologic. Wherever other lithologies were exposed at the time of passage of the ash flows, lithic fragment assemblages in the tuff become more diverse, as seen in the eastern Paulina Basin (Fig. I-4).

Because the tuff normally contains very few lithic fragments, grading characteristics are difficult to evaluate. Lithic clast size generally decreases with distance from the vent, based on the distribution of lithic fragments where four to five measurements were taken per locality and where entrainment of local large clasts was not an obvious problem (Fig. I-7b). The correlation is weak ($R^2 = 0.15$). At several localities, pronounced lithic-enriched zones formed within the basal meter of the tuff. A local derivation of lithic fragments for these enriched zones is indicated by their similarity with substrate lithologies, by ramp-like alignments of lithics from the substrate into the tuff upward at an angle of ca. 20° (Fig.I-8) (similar pick-up zones are described by DeSilva, 1989), and by the abnormally high average size of lithics (Fig. I-7b). Anomalously large lithic fragments, thought to be locally derived, are also described from the Miocene Peach Springs Tuff in the southwestern US. (Valentine et al., 1989). Local derivation of lithic fragments and lithic breccias of the Peach Springs Tuff is attributed to surface roughness that may have induced a turbulent boundary layer capable of picking up loose substrate material which was soon deposited (Buesch, 1992).
Speculations on pumice and lithics variations

In ignimbrites, average lithic fragment size typically correlates better with distance than average pumice size (Sparks, 1975, Cas & Wright, 1987, p.193, Suzuki-Kamata & Kamata, 1990). Lithic fragments are more likely to be sorted with distance than are pumices because they are significantly denser than the ash flow, whereas pumices may be slightly more or less dense than the ash flow (i.e., reverse vertical grading of pumices and normal vertical grading of lithic fragments Cas & Wright, 1987, p.189). Therefore, the controlling factors in introducing distance dependencies are: (a) size and density of clast, (b) density or viscosity of the ash-flow (Cas & Wright, 1987, p.193).

In the Rattlesnake ignimbrite, the size-distance correlation is quite strong, implying that the matrix of the ash flow was expanded to a degree that allowed pumices to settle out. The ungraded vertical pumice distribution, however, suggests it was not a strongly expanded flow. Experiments indicate that more expanded flows develop stronger reverse grading in pumices (Wilson, 1980). One could explain this apparent paradox with a model where material separated out of overriding highly expanded pyroclastic flows to produce denser, near surface ash flows which either did not travel far enough or were sufficiently dense to prevent pumice clasts from becoming reversely graded. For the Campanian ignimbrite, Fisher et al. (1993) proposed a similar model, in which the transport medium consisted of extremely expanded ash flows in order to travel over several hundred meter high ridges and the deposition medium consisted of near surface, locally developed, much less expanded ash flows. The exponential decrease in average maximum pumice clasts away from the source in the Rattlesnake Tuff is analogous to exponential pumice size decrease commonly observed in subaerial fall out tephra (Fisher & Schmincke, p. 151, 1984) also indicating that a highly expanded "transport medium" existed which exerted principal control on the pumice distribution in the Rattlesnake Tuff.

The poor correlation of lithic fragments with distance can be interpreted to mean that at no point were the ash flows capable of transporting large lithics over long distances. Younger basalt lavas cover the tuff within 20 km of the proposed venting area, so the strongest (near-vent) lithics-distance correlations are undocumented. Wright & Walker (1977) and Suzuki-Kamata & Kamata (1990) observed most rapid changes with distances between ca. 5 - 15 km. The interpretation of lithic fragment distribution is complicated by local entrainment of clasts from the ground in areas of greater basal erosion consistent with increased turbulences around topographic obstacles (c.f. Buesch, 1992) or around areas where interaction with water can be inferred. Deposition of the lithic clasts occurred soon after entrainment, forming sometimes pronounced lithic-enriched zones at the base of the tuff. Rattlesnake Tuff from the John Day Basin (Fig. I-4) has a zone enriched in rounded lithics that match conglomerate substrate lithologies. There, evidence for possible involvement of water is found nearby in degassing pipes (Enlows, 1972: Fig.9), secondary ash flows (both found nowhere else), and stronger columnar jointing. The main conclusion is that the average size distribution of lithic fragments with distance only partially reflects the continued settling of increasingly smaller lithics due to lower suspension capabilities of the flows, but it also reflects the changing erosion potential which is a product of conditions imposed by the topography, substrate, and the energy of the flow.

Source area

No caldera structure related to the Rattlesnake Tuff is exposed. By analogy to pyroclastic deposits of similar volume, one might expect a caldera with an approximate diameter of 20 km (c.f. Smith, 1979, Spera & Crisp, 1981). The proposed vent area is covered, so whether or not the eruption produced a caldera cannot be resolved at present.

Several source areas, all lying within the Harney Basin, have been proposed (Walker, 1969, 1979, Parker, 1974, MacLeod et al., 1975). One of the first proposed source areas was in the Buzzard Creek area (Fig. I-1 and I-10) (Walker, 1969, Parker, 1974) as deduced from rheomorphic features interpreted as venting fissures (Walker, 1969). Later, a caldera now buried under Harney Lake was proposed as source area (Walker, 1970, 1979). A venting area located in the western Harney Basin was favored based on a clustering of silicic domes of similar age (MacLeod et al., 1976).

In order to evaluate the potential source area of the Rattlesnake Tuff, we used an internally consistent model of increasing pumice size toward the vent and compared the proposed source to the areal distribution, distribution of facies, and flow-direction indicators in the tuff. The pumice data locations were cast in terms of a polar coordinate system with the origin as the presumed source. By relocating the origin of this coordinate system for different proposed venting areas with recalculation of the polar coordinates to the same pumice data set, one can evaluate how well the average pumice size and distance correlate for different source areas. Exponential regression lines were calculated to see which source area would give the best fit to the available data.

Fig. I-8 Ramp-like lithics zone

Alignment of lithic fragments from the substrate into the tuff is thought to represent a picking-up feature at a site 66 km from inferred source, note hammer among largest lithics.



Fig. I-9 Source evaluation

Figure shows exponential fit to recalculated pumice data to different source areas (point source). Linear regression fit gave a poorer correlation for each site. Refer to figure I-10 for source area locations. Capehart Lake is the proposed venting site shown in other figures.



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Fig. I-10 Pumice alignments and source areas

Average alignments of long axes of pumices; arrows indicate directions of average alignments. Source area locations of evaluated sites from figure I-9 are shown with stippled ovals, Capehart Lake site is inferred source shown in all other figures.



The best fits are for a source area near "Capehart Lake" located in the western Harney Basin, 18 km south-southwest of Riley or 30 km to the west of "Capehart Lake" (Fig. I-9). This finding is compatible with measured average alignments of the long axes of pumices (Fig. I-10). We prefer the Capehart Lake source area because it lies in the middle of the distribution of strongest crystallization and rheomorphic features of the Rattlesnake Tuff (Fig. I-16, I-17) and within the arcs of thickest tuff (Fig. I-5). Moreover the proposed vent is near the center of the tuff's areal distribution (Fig. I-4). It is almost identical to the source area proposed by MacLeod et al. (1976), based on the regional age distribution pattern of volcanism (Fig. I-1).

Welding and crystallization facies

Terminology, appearance and occurrence

Most of the welding and crystallization terminology used is according to Ross & Smith (1960) and Smith (1960a, b). With respect to spherulites and lithophysae, Iddings' (1885-86) treatment was consulted. Because the Rattlesnake Tuff displays such a variety of welding and crystallization facies, the attempt was made to differentiate facies in more detail when mappable.

Welding facies

Five welding degrees are distinguished which occur in vitric tuff. From least to most welded, the degrees are: nonwelded, incipiently welded (or sintered), partially welded with pumice, partially welded with fiamme, and densely welded (or strongly welded). Transitions between welding degrees are highly gradational.

In the nonwelded tuff, no to very little adhesion among glass shards exists. Glass shards and pumice clasts have kept their original shape. In the Rattlesnake Tuff, nonwelded tuff is only found at the base of sections. Thicknesses of lower nonwelded tuff range from about 0.5 m in near source, highly zoned sections to 4-6 m in thick distal sections. The specific gravity is 1.43 g-cm⁻³. In incipiently welded (or sintered) tuff, some adhesion among glass shards is developed without destroying the original textural features of shards and pumices (Fig. I-3a & b); no coalescence of glass material has occurred. Three specimens yielded specific gravities of 1.48, 1.57, 1.64 g-cm⁻³, with an average of 1.56 g-cm⁻³. Unlike the nonwelded tuff, sintered tuff forms rugged cliffs. Erosion characteristics of the sintered facies are distinctive; meter-size boulders break off exposed cliffs but continued erosion takes place in the form of loosening single shards or pumices from the exposed surface of the tuff rather than forming smaller cobbles. Handsamples look dull, the glassy nature is seen with the aid of a handlens. In highly zoned sections, sintered tuff occurs as a thin, poorly defined layer, tens of cm thick, within the lower vitric part. Sintered tuff is best developed in sections which are sintered throughout, reaching thicknesses of up to 20 m.

The zone of partial welding of Smith (1960a, b) was divided into a lesser degree of partial welding, the "partially welded with pumice", and a stronger degree, the "partially welded with fiamme". The distinction is mappable and becomes most important in intermediate to distal portions of the tuff (see regional facies).

The welding degree "partially welded with pumice" is characterized by slight deformation of glass shards and pumices. Hand specimens classified with this welding degree yielded specific gravities of 1.63, 1.67, 1.74, 1.80, and 1.83 g-cm⁻³. Hand samples have a glassier luster than the incipiently welded degree, however, pumices embedded in the matrix have not yet collapsed to form fiamme; originally round bubble shards are only slightly ellipsoidal under the microscope. No recognizable darkening is associated with this welding increase. Weathering produces more platy and angular forms; fragmentation ranges from boulders to pebbles. In highly zoned sections, this zone is found in the lower and upper vitric parts. In some sections with thicknesses as great as 14 m, the main central part consists of this welding degree. In upper vitric tuff, there is a color change from gray to reddish due to primary oxidation by upward percolation of vapors during cooling of the tuff.

The welding degree "partially welded with fiamme" is reached when most of the pumices have are flattened to fiamme, while the overall appearance is darker and glassier, but does not yet appear as black vitrophyre (Fig. I-11). Specific gravities for such specimens are 2.15, 2.20, and 2.29 g-cm⁻³. Higher densities reflect the continued decrease in pore space. Samples transitional between partially welded facies, that is, having both pumice and fiamme, have intermediate densities; four specimens yielded 2.07, 2.11, 2.12, 2.13 g-cm⁻³. The occurrence of "partially welded with fiamme" ranges from several meters thick in the central portion of sections at greater distances (>

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60 km) from the source to a thin (ca. 10 to 50 cm) zone marking the transition to the densely welded tuff within the lower vitric part of strongly zoned sections.

The highest welding degree results in densely welded (or strongly welded) tuff that is an obsidian-like, black vitrophyre. The highest specific gravity is 2.34 g-cm⁻³. Densely welded tuff occurs within a radius of ca. 80 km around the inferred vent. The glassy, densely welded zone is thickest (ca. 4 m) where it is overlain by a pervasively devitrified zone. The zone thins to 1 m where it is overlain by a spherulite-lithophysal zone. Therefore, thickness of the vitric, densely welded zone negatively correlates with stronger degree of crystallization in the overlying zones (see below).

In summary, welding degrees range from non- to densely welded with associated specific gravities showing a continuum from around 1.43 to 2.34 g-cm⁻³; this range corresponds well with specific gravities of other tuffs (e.g. Martin, 1959, Ross and Smith, 1960, Smith, 1960b, p.46, Ragan & Sheridan, 1972). As based on Rattlesnake Tuff samples, the preferred division of the density spectrum into welding degree is:

< 1.5 g-cm ⁻³	=	nonwelded;
1.50-1.65 g-cm ⁻³	=	incipiently welded;
1.65-2.05 g-cm ⁻³	=	partially welded with pumice;
2.05-2.30 g-cm ⁻³	=	partially welded with fiamme; and
2.30-2.34 g-cm ⁻³	=	densely welded

Noteworthy is that the partially welded tuff with pumice and the partially welded tuff with fiamme are not very different in appearance compared to the black vitrophyre, yet they span the greatest density change from around 1.7 to 2.25 g-cm⁻³. On the other hand, the transition to a black vitrophyre is hardly associated with an increase in specific gravity. Indeed, one hand specimen with indistinguishable grayish fiamme and matrix yielded the same density as black vitrophyre. This gray vitrophyre represents vitric tuff observed in a few outcrops in the northern Paulina Basin and areas to the east (Fig. I-4).

Fig. I-11 "Partially welded with fiamme" welding degree

Most pumices have compacted to fiamme, while groundmass is middle gray. Density of specimen is 2.29 g-cm⁻³, coin is 2.4 cm in diameter.



Fig. I-12 Vapor phase crystallization

Top: Cryptocrystalline vapor-phase facies. Vitroclastic texture is retained despite 100% devitrification into very fine elongated crystals forming axiolitic structures, under plane light; white areas are vesicles. Field of view is ca. 3 mm.

Bottom: Microcrystalline vapor phase facies. Same scale as above with crossed nicols, crude vitroclastic texture seen only under microscope due to coarseness of more granular crystals. Note vesicles (black ovals) lined with largest crystals.

Fig. I-12:



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Fig. I-13 Vitric - vapor phase transition

Cryptocrystalline vapor-phase zone overlying vitric tuff. Sharp interface (at level of hammer) separates a lower vitric, partially welded tuff with pumice from an upper cryptocrystalline vapor-phase zone where holes indicate dissolution of pumice clasts. Site is 63 km from inferred vent. Similar vapor-phase tuff is seen in figure I-12a.



Crystallization facies

Primary devitrification processes led to the formation of four crystallization zones, the vapor-phase zone, the pervasively devitrified zone (identical to the devitrified zone of Smith, 1960a), the spherulite zone, and the lithophysae zone. Transitions are mainly gradational between crystallization zones and vary in thickness from around 0.3 m to as much as 2 m.

According to the grainsize of the crystals formed during the crystallization process, a cryptocrystalline and a microcrystalline vapor-phase zone are distinguished (Fig. I-12). The two subzones can be distinguished in handspecimen with a handlens. In the cryptocrystalline zone, the vitroclastic texture is perfectly preserved combined with partial mineral precipitation in open pore space. No glassy material is present, but single groundmass crystals are not recognizable. The very fine axiolitic structure of the shards is revealed under the microscope (Fig. I-12a); the width of single crystals is < 1 micron. In contrast, in the microcrystalline zone the axiolitic structures are coarser grained with a crystal grainsize of ca. 5 to 150 microns; largest crystal growth occurs along vesicle walls (Fig. I-12b), showing the zone-characteristic crystallization into open pore space (Smith, 1960a, p.155). Despite the larger grain size, the original forms of the shards are retained even if not recognizable in handspecimen where single crystals are discernible. Both zones are developed in tuff that is sintered to partially welded with pumice. The transition between the two vapor-phase varieties is difficult to place. In proximal sections (< 40 km) where the vapor phase zone directly overlies the basal vitric zone (Fig. I-13), the microcrystalline variety seems to develop in the upper part of the vapor phase zone. Only microcrystalline vapor-phase tuff was observed to overlie pervasively devitrified tuff. Thicknesses of the vapor-phase zone range from 8-10 m, where overlying vitric tuff, to < 0.5 m, where sandwiched between pervasively devitrified and upper vitric tuff of highly zoned sections.

The *pervasively devitrified crystallization zone* is developed in tuff that is densely welded or partially welded with fiamme, and is clearly recognized by the stony appearance. It can also be distinguished from the vapor-phase tuff by its greater density. There is good retention of the flattened shards structure, as observed in thin sections. Crystal sizes are similar to the microcrystalline vapor phase tuff. Thickness ranges from 15 m, where directly overlying black vitrophyre, to 2-4 m, where overlying lithophysal tuff.

Fig. I-14 Lithophysae in devitrified tuff

Top: "Lithophysae in devitrified tuff" facies. Lithophysae show continuum from solid to completely hollow, matrix material is dense stony devitrified tuff. Note larger gas cavity along bottom of picture. Coin for scale is 2.4 cm.

Bottom: Close up of beginning lithophysa in pervasively devitrified tuff. Lithophysa rind surrounds devitrified tuff. Note ghost shards between lithophysa rind and top of lower lithophysa. Field of view is ca. 6 mm.



Fig. I-15 Rheomorphic tuff

Small scale folds in rheomorphic tuff. Pen is 14 cm in length.



The most complexly distributed crystallization zones are the *spherulite zone* and the *lithophysae zone*.. Spherulites range in size from 1 mm to about 2 cm and develop almost exclusively in strongly welded vitric tuff right below the abrupt vitric/devitrified boundary (see transitions). Only a few sections were observed where spherulites formed without lithophysae.

Unlike the other crystallization zones lithophysae growth is an overprint on both vitric and devitrified tuff. According to the matrix material in which lithophysae were formed, three zones occur that make a consistent stratigraphic sequence: the lower zone of lithophysae in perlitic black vitrophyre, the middle zone of lithophysae in spherulites, and an upper zone of lithophysae in devitrified tuff (Fig. I-14). Beeson (1969) and Walker (1979, 1970) described parts of the above sequence. A gradational transition is found between lithophysae in vitrophyre and lithophysae in spherulites, whereas the second transition to lithophysae in devitrified tuff is abrupt. The very subtle interface separating lithophysae in spherulites and lithophysae in devitrified tuff is likely to represent the vitric/devitrified boundary before lithophysae developed. The lower zone, lithophysae in vitrophyre zone, is typically 1-2 m thick. The average thickness for the zone lithophysae in spherulite tuff is 2-3 m. The zone of lithophysae in devitrified tuff is thickest, up to about 40 m, however in one section this zone was only ca. 1 m.

The lithophysae are 1-3 cm in diameter. The abundance within the lower zones increases upward from a few dispersed lithophysae in densely welded vitric tuff to a network of lithophysae in spherulitic matrix. Lithophysae in the lower two zones are mainly solid with minor development of hollow forms (Iddings, 1885-86, Pl. XII and XIV). The lithophysae in devitrified matrix are similar in size and are typically equidimensional (Fig. I-14) but, may be all solid or completely hollow with a crystallization rind lining the walls. Transitions include lithophysae with a hollow part or loose interior pellet.

Based on thin section and slab studies, lithophysae observed in devitrified matrix nucleated not in individual spots from where they grew outward, but several examples demonstrate that the first part of a lithophysa is a crystallization rind surrounding some pervasively devitrified tuff. This narrow rind grows mostly outward to form filled lithophysae which, with continued crystallization, may become partially and then completely hollow (Fig. I-14). The loose interior pellet in hollow lithophysae represents the remaining core of pervasively devitrified tuff which was first surrounded by lithophysae crystallization. By analogy, lithophysae formed in pervasively devitrified and spherulitic tuff are therefore mainly a recrystallization phenomenon. There is no vapor phase crystallization associated with the lithophysae. The generation of hollow lithophysae solely through release of volatiles as proposed by Ross and Smith (1960, p. 38) seems questionable in light of the documented evolution from filled to hollow lithophysae. Therefore we distinguish between lithophysal cavities and gas cavities.

Previously, lithophysal and gas cavities were often used synonomously (e. g. Smith, 1960a, p. 156). Despite both being cavities, we suggest their origin might be different. Gas cavities are the result of entrapment of gases (e. g. Smith, 1960a) in contrast to lithophysal cavities (or hollow lithophysae) that are crystallization phenomena. In gas cavities, vapor phase crystallization might deposit crystals along the walls (Smith, 1960a, p. 156); the characteristics of vapor phase minerals should help to distinguish them from hollow lithophysae.

In non-rheomorphic parts of the Rattlesnake Tuff, gas cavities are always completely hollow, round to ellipsoidal, often rugged in outline, and are not associated with vapor phase minerals. Near the source, the largest gas cavities (10-40 cm) are common in the upper part of the lithophysae zone. In thick distal sections, gas cavities are small (ca. 4 cm) and elongate and occur in the absence of lithophysae or lithophysal cavities (see regional variations).

Rheomorphic tuff

The Rattlesnake Tuff has pronounced secondary flowage features (e.g. Wolff & Wright, 1981) at numerous places within 50 km of the vent (Fig. I-15) (see regional facies). The dominant appearance of rheomorphic Rattlesnake Tuff is strongly lineated devitrified tuff with locally abundant elongate openings parallel to foliation. Whether the majority of these openings originated as gas cavities and were stretched later or whether they formed by separation of flow layers cannot be answered although the extent of some of them suggests the latter. In some thick, devitrified rheomorphic units these openings are lined with vapor phase minerals. Rheomorphic tuff may also be glassy, ranging from densely to nonwelded. Most dense, vitric rheomorphic tuff has the black color and vitric appearance of a densely welded vitrophyre plus typical elongated flow features, such as lineations. The most extraordinary rheomorphic tuff consists of a loose ashy rheomorphic matrix with flame-like structures that likely represent deformed pumices. There is almost no cohesion of the ash-material. Only one occurrence on top of an entirely rheomorphic section could be documented.

In principle, there are two types of rheomorphic sections according to the gradation upward from the central devitrified rheomorphic part. Devitrified rheomorphic tuff grades into normal tuff in one case and into rheomorphic vitric tuff in the other case. Both types have a normal developed non-to densely welded, vitric base. Above this normal base, the first rheomorphic tuff is either vitric or devitrified.

Transitions between facies

The transition from one welding or crystallization facies to another can be smooth, gradational, or abrupt. Transitional zones are typically more abrupt the more zones that are developed. The most abrupt transitions are between lower vitric and devitrified zones and between lower vitric and vapor phase zones (Fig. I-13). These transitions always appear as a pronounced interface 2-5 cm thick. Sharp planar and subhorizontal contacts between glassy and devitrified zones are also reported from high-temperature ash-flows tuffs in Idaho (Bonnichsen & Citron, 1982, Bonnichsen & Kauffmann, 1987). Similar, but not as pronounced, are the transitions between devitrified zone is obliterated, but still traceable after the overprint of spherulites and lithophysae. Transitions from normal to rheomorphic tuff are always gradational over 30 cm to more than one meter. All contacts within non-rheomorphic tuff are in most cases nearly horizontal. Exceptions are found at a few places where the sharper transitions are inclined as much as 20^o (see local facies variations).

Facies model

In this section, we address variations in the facies distribution which become apparent by comparing local facies variations to those of areas at different distances from the source. If not otherwise stated, the variations are independent of tuff thickness because thicknesses of nearly complete sections are more or less constant (Fig. I-5). In general, the greatest diversity of facies is found closest to the source, where all facies occur, and decreases outward. Results are summarized in figures I-16, I-17, and I-18. Combining zonal variations of 85 individual sections with regional facies distribution patterns produces an overall facies model that documents the vertical and horizontal variations in welding and crystallization characteristics in the Rattlesnake Tuff (Fig. I-18). In figure I-18, the axis "thinner" is the axis which describes local variations; regional variations are displayed by four intervals at increasing distance from the proposed source area. The distance intervals are based on the approximate distances where changes become noticeable (see regional variations). The principal variations in the Rattlesnake Tuff are represented by the "common sections" (Fig. I-18) and describe most of the tuff. Features of tuff outcrops that are unusually thick or outcrops displaying rheomorphic flow feature are illustrated with separate block diagrams as "thick and rheomorphic sections". Upper nonwelded tuff was encountered only once, and thickness variation would be purely speculative; it was therefore not incorporated into the facies reconstruction.

Regional facies variations

The degree of welding decreases with distance from the source. To generalize the distribution of welding facies approximate boundaries were drawn to indicate changes in the highest welding degree developed (Fig. I-16). The greatest change takes place between 60 to 80 km from the source (Fig. I-16). Beyond a distance of ca. 80 km, densely welded tuff (black vitrophyre) is rare. Instead, partially welded tuff with fiamme becomes the highest welding degree. The change to where partially welded tuff with pumice is the strongest welding degree probably lies around 130 km.

Envelopes around the most distal outcrops of rheomorphic, lithophysal, vapor phase (without underlying devitrified tuff) and devitrified tuff are shown in figure I-17. At a distance of ca. 80 km, the occurrence of lithophysal tuff and of vapor phase zones decreases. Lithophysae and lithophysal cavities beyond 80 km are extremely rare. At distances greater than 80 km, the occurrence of vapor phase tuff is minor and only above a lower devitrified zone. Closer to the vent, vapor phase alteration zones also occur frequently directly above vitric tuff, reaching maximal thicknesses of ca. 10 m (Fig. I-13). Vapor phase zone overlying vitric tuff without any pervasive devitrified tuff in between proximal to the venting site has been undocumented in the literature.

Rheomorphic tuff localities are restricted to an area within 40-60 km of the proposed vent. Most rheomorphic tuff localities coincide with thickest tuff which in turn occurs as 10 to 20 km wide arcuate belts (Fig. I-5).

Fig. I-16 Distribution of welding

Regional distribution of welding facies, with respect to highest welding degree observed at a given locality. Rattlesnake Tuff outcrops are shown with closely stippled pattern. Distribution of highest welding facies is more interpretive in areas with thin outcrops (< 5 m). Insert shows area of figure I-16 relative to Oregon.





Fig I-17 Distribution of crystallization

Regional distribution of crystallization facies. Arrows with letters indicate areas mentioned in "Local Facies Variations", viz., LL: Lunch Lake, SC: Silver Creek, FH: Fort Harney, TB: Twin Buttes.





Fig. I-18 Facies model

Three-dimensional facies model. Areas without pattern are vitric. Pre-erosional upper nonwelded tuff omitted, except in one section displaying a characteristic rheomorphic section. Devitrified zone is synonymous to pervasively devitrified zone. Thicknesses indicate range of observed thicknesses of least eroded sections.



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Fig. I-18:

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Local Facies variations

At several localities, strong variations in welding and crystallization facies occur over a distance of a few hundred meters with little or no change in thickness. Strong local topographic variations can be excluded as a cause of these facies variations in most cases. The detailed mapping of these facies variations is the basis for the model presented in figure I-18. Four areas, at different distances from the inferred source are described below from the closest to the most distal locality.

Silver Creek and Lunch Lake are located at the same distance from the inferred source (31 and 33 km, respectively) (Fig. I-17). In both instances, the facies variations over a distance of 1-2 km exhibit the transition from a completely incipiently welded, vitric section to a highly zoned section with a lower non- to densely welded vitric zone overlain by a crystallized zone of dominantly lithophysal tuff. Sections transitional between the endmembers include sections with lower partially welded tuff with pumice or with fiamme, both overlain by a vapor phase zone and sections where non- to densely welded vitric tuff is overlain by pervasively devitrified tuff without any lithophysae. In the Lunch Lake area the described transitions take place along a flatlying, ca. 15 m thick, continuously exposed Rattlesnake Tuff rim which conformably overlies a basalt flow. There within 200 m of continuous outcrop, the tuff rim changes from lower partially welded vitric tuff with fiamme overlain by cryptocrystalline vapor phase zone, to a section with densely welded vitric tuff overlain by pervasively devitrified tuff which is capped by microcrystalline vapor phase zone and, finally, to a section with a several-meter-thick core of lithophysal tuff underlain by densely welded vitric and overlain by pervasively devitrified tuff. The described changes are not due to distance from the inferred vent because, in both areas, the completely incipiently welded section is closer to the vent than any of the more welded and more crystallized section.

At Fort Harney, 63 km from the inferred source, Rattlesnake Tuff is 10-20 m thick. Despite the greater distance, facies variation from an incipiently welded tuff to a section dominated by lithophysae also occurs over 2 to 3 km. A 100-m-long outcrop of tuff nicely displays how vitric, partially welded section with pumice is split into lower and upper vitric tuff separated by a central wedge of cryptocrystalline vapor phase tuff.

At Twin Buttes, 83 km from the inferred source (Fig. I-17), the lateral gradation in facies occurs over 50 m and the change is from a 15-m-thick (substrate to tuff is exposed), entirely vitric section, composed mainly of partially welded tuff with pumice, to tuff with an upper 7-m-thick pervasively devitrified zone and a vitric,

partially welded zone with fiamme at the base. A sharp interface, marking the vitric/crystallized transition, appears where the first crystallized tuff makes up the top of the section. From there, the interface runs towards thicker crystallized tuff downward at an angle of 25^o degrees relative to the horizontal tuff outcrop.

Discussion of welding and crystallization facies

Comparison to existing facies models

Previously published facies models for ignimbrites are largely based on idealized hypothetical models established by Smith (1960a). His models were constructed combining data from a large number of ash-flow deposits to show general differences in the distribution of welding and crystallization zones in end-member type ash-flow tuffs. Three end-member facies models include (1) a deposit which is thick, but too cold to develop strong welding and crystallization zones, (2) a thin, hot deposit which developed stronger welding zones, but without much crystallization, and (3) a deposit which was sufficiently thick, hot, and gas rich to develop dense welding and crystallization zones (Smith, 1960a, plate 20). All of the displayed variations within each model are attributed to a decrease in temperature and thickness of the deposit with increased distance from the vent even though some deposits indicate only very slight effects of cooling or thinning with distance (Smith, 1960a). In comparison, the zonal variations in the Rattlesnake Tuff (Fig. I-18) document the complexity of a single ashflow sheet that is about constant in thickness; strong facies variations occur extremely locally. Thickness and distance have only subtle correlation with welding and crystallization changes.

Timing

The relative time sequence of the welding and crystallization processes can be deduced from textural and field evidence. Welding is the first process to begin. Any noticeable adhesion of glass shards during the last stages of flow, known as primary welding (Chapin and Lowell, 1979) or as agglutination (e.g. Branney & Kokelaar, 1992), is excluded because entirely incipiently welded sections also occur closest to the vent. Therefore, welding phenomena in the Rattlesnake Tuff are post-depositional.

Crystallization followed welding because it is superimposed on welding textures and some crystallization facies are associated with particular welding facies. The vapor phase zone occurs in partially welded tuff. At higher welding degree, the pervasive devitrified zone develops instead. Individual spherulites are found only in densely welded vitric tuff. Where spherulite growth led to an interconnecting network of spherulites, the previous nature of the tuff is obliterated but it appears that spherulite growth is restricted to densely welded, originally vitric tuff. Therefore, spherulite formation can be placed after welding and is likely to postdate the main formation of the pervasive devitrified zone (see below). Evidence from silicic lava flows also suggests that formation of spherulites occurs in glassy material remaining after the central parts devitrify (Bonnichsen & Kauffmann, 1987). The last crystallization process to occur is the development of lithophysae which overprints vitric, spherulitic, and pervasively devitrified tuff. That lithophysae in the Rattlesnake Tuff formed after and not during pervasive devitrification is based on the observation that early stages of lithophysae are crystallization rinds that destroy the axiolitic devitrification of still recognizable shards. Lithophysae formation through recrystallization does not only occur in the pervasive devitrified zone but also in the spherulite zone where spherulites are overprinted. The subtle and diffuse interface recognized in lithophysal tuff, separating spherulitic from pervasively devitrified tuff, likely represents the original sharp interface separating vitric from devitrified tuff, suggesting that pervasive devitrification is followed by later spherulite formation in some remaining glass, and crystallization of lithophysae is last.

Concerning the timing of rheomorphism, it cannot be excluded that upper parts of strongly draping tuff developed its foliation through a continuum of primary to secondary processes; however, the main phase of flowage is thought to be of secondary origin where the basal section is normally developed and where strong folding with variable orientation suggests no strong substrate relief.

Causes of facies variations

Especially, the drastic facies variations over <1 km to 2 km raise the question of their cause. A number of factors influence welding and crystallization of which the most important are temperature, pressure, volatiles, and composition (Ross and Smith, 1980, Smith, 1960a, b). Each factor will be discussed in the sequence of greater

importance which is deduced from considerations concerning primarily local facies variations in the Rattlesnake Tuff.

Differences in bulk magmatic composition are regionally negligible and are most likely locally constant because 99% or more of tuff is high-silica rhyolite. There are no local concentrations of dacitic pumices.

The influence of lithostatic load on the distribution of welding is thought to be very minor because of the relatively uniform thickness of the tuff. Differences in specific gravity indicate densely welded tuff was never more than 1.7 times as thick as nonwelded tuff. Also, the missing nonwelded top was likely not markedly thicker over densely welded versus nonwelded sections, based on analogy to preserved basal nonwelded zones. Recently, a similar interpretation was reached for the Sifon ignimbrite, northern Chile, where degree of welding is apparently unrelated to thickness (DeSilva, 1989).

Volatiles can be divided into magmatic volatiles and inherited volatiles. Magmatic volatiles released from vesicles and from continued degassing of glass are likely to decrease with travel distance from the vent. However, it seems highly unlikely that magmatic volatile composition or contents varied significantly on a local scale. On the other hand, volatiles derived from the substrate may change over short distances. The degree of welding along the base in modeled ash-flow sheets is drastically reduced where emplacement in shallow water occurred (Riehle, 1973). Therefore, the expected influence of enhanced water influx from a wet substrate would be chilling of lower parts of the Rattlesnake tuff inhibiting strong welding. Local welding variations may record differential water influx, however it does not explain or rather contradicts the accompanied strong variations in crystallization facies assuming stronger crystallization is also caused by the increased influx of volatiles from the substrate. Consequently, stronger crystallization should be accompanied by a lesser degree of welding in the lower vitric zones. The opposite is observed.

Subtle changes in the temperature conditions of the tuff seem the likeliest cause of the abrupt facies variations. The temperature the tuff immediately after emplacement depends on the initial magmatic temperature and the heat lost during emplacement which in turn depends on accumulation rate and cooling rate. Accumulation and cooling rate depend on the emplacement mechanism. The high-energy character of the Rattlesnake Tuff and the lack of major topographic lows where thick sections could accumulate indicate a very similar emplacement mechanism locally. Therefore, we argue that regardless of the complex interplay of these factors, they are not likely to cause large differences in temperatures right after emplacement over the short distances

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characteristic of local facies changes. We propose that the post-emplacement temperature history, the cooling to ambient temperature, might have been different enough to control local facies variations. Welding and crystallization could be controlled by a step-function behavior and occurs when a tuff reaches and stays at a critical temperature for some amount of time. Thus fairly subtle differences in thickness (thermal insulation) could have important effects.

Conclusions

The Rattlesnake Tuff was deposited from high-energy ash-flows emplaced as multiple flow units. Pumice distribution and grading characteristics favor the hypothesis that emplacement was separated into a transport and deposition medium. Emplacement was rapid enough to develop a single cooling unit with vertical and lateral variations in welding and crystallization facies. The tuff erupted from an area near the center of today's outcrop distribution, based on pumice clast distribution and the regional variation of welding and crystallization facies. Vitric welding facies range from nonwelded to densely welded with several intermediate welding degrees, namely incipiently welded, partially welded with pumice, and partially welded with fiamme. Early crystallization overprints welding. Distribution of vapor phase alteration and pervasive devitrification is governed by previously developed welding degrees. Among vapor phase alteration, a cryptocrystalline and microcrystalline facies are distinguished as based on crystal size. Late crystallization features include spherulites growing in densely welded vitric tuff and lithophysae which overprint vitric and all other crystallization facies, except possibly vapor phase tuff. Within 50 km around the source area, secondary flowage processes generated vitric to devitrified rheomorphic tuff.

Strong local variations, over <1 km to 2 km, include the complete spectrum of regional variations, exhibited over 10's to 100's of km's. Local facies variations are thought to be caused mainly by slight variations in original thickness near the critical temperature and volatile content required for welding and crystallization.

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Chapter II: The Rattlesnake Tuff, part I: Relationships between high-silica rhyolites, dacites and mafic inclusions

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Abstract

The Rattlesnake Ash Flow Tuff from eastern Oregon represents ca. 280 km³ of high-silica rhyolite magma erupted as pumices and glass shards. Dacite pumices make up less than 1% of the total volume and quenched basalt and basaltic andesite inclusions inside dacite pumices constitute << 0.1% volume percent.

Trace and major element variations among high-silica rhyolite pumices indicate a series of progressively more evolved compositions. Derivation of least-evolved highsilica rhyolites through dehydration melting events is the process which is most compatible with the chemical record. Major element composition of least evolved Rattlesnake Tuff high-silica rhyolites and melts obtained from dehydration experiments are similar. Ba/Rb ratios of 30 ± 5 and La_N/Yb_N of 4.5 ± 0.2 for the group of least evolved high-silica rhyolites (group-E pumices) constrain potential protolith compositions to some amphibolites, high-Ba greywackes or granulite facies intermediate to mafic rocks.

At least three types of mafic inclusions, ranging in size from cm to mm, are recognized which are mainly found inside dacite or dacite/rhyolite banded pumices. Ubiquitous glomerocrysts of plagioclase and chrysolite characterize the inclusion type which is similar to regional high alumina olivine tholeiite (HAOT) lava flows typical of the Oregon Plateau. Phenocrysts and groundmass show strong quenching features. Phenocryst-poor, basaltic andesite inclusions with a micro-quenched groundmass is the second type. Such inclusions have round to streaky forms often with mingled textures with the host pumice. Phenocryst-poor inclusions acquired their enriched trace element signatures mainly through fractionation and recharge processes. The third inclusion type is also basaltic but consists of clinopyroxene which poikilitically encloses plagioclase phenocrysts or of granular olivine with plagioclase. Dacite pumices (62-68 weight % SiO2) are phenocryst poor with 1 to 4 % crystals. Mineral assemblages are strongly bimodal with euhedral, resorbed, or reacted phenocrysts from high-silica rhyolites or from basaltic magmas. Dacite magmas are likely to have been generated through mixing of least evolved high-silica rhyolites with enriched basaltic andesite magma represented in phenocryst-poor basaltic andesite inclusions. A silica gap of ca. 6 weight % exists between dacites and high-silica rhyolites. Only pumices with strong banding fall in the gap.

The reconstructed magma chamber was stratified from high-silica rhyolites at the top to mafic magmas with dacites at the interface. The compositional range of the pre-eruption magma chamber is based on banded pumices containing all compositions. The stratification is inferred from vertical and lateral distribution of erupted compositions within the Rattlesnake Tuff. This stratification is likely to have been stable over some time, based on a petrologic model which explains the relationships between high-silica rhyolites, dacites, and basaltic compositions. After formation, a silicic magma chamber impedes mafic magmas (HAOT) to ascent therefore causing the mafic magma to fractionate (enriched mafic inclusions) which in turn thermally stabilized the silicic magma causing it to fractionate (zoned high-silica rhyolites). Mafic and silicic melts interact and form dacites along their interface.

Introduction

Compositional zonation of magma systems as recorded by zoned ash-flow tuffs is common and generally of two types (e.g. Smith, 1979, Hildreth, 1981): (1) zonations within rhyolites in which trace-element vary widely with only minor ranges in major element concentrations; and (2) zonation with a wide range of compositions including silicic chamber tops and more mafic underpinnings (Smith, 1979, Hildreth, 1981). The first type of zoning is typical of large volume tuffs(> 100 km³) and the second is more commonly observed in small volume tuffs ($< 50 \text{ km}^3$). The Rattlesnake Tuff is of large volume (280 km³ dense rock equivalent), forms a single cooling unit, and consists dominantly of high-silica rhyolite strongly zoned in trace elements but also includes abundant evidence for a dacite & basaltic root zone. The large volume combined with both types of zonation and excellent preservation of pumices make the Rattlesnake Tuff uniquely suited to investigate the compositional variety and interaction of magma types.

Fig. II-1a Regional setting

Outcrop pattern of Rattlesnake Tuff (dark gray) with respect to Oregon; light stipple represents Blue Mountains Province, dense stipple outlines High Lava Plains. Dotted lines with numbers are simplified isochrons in million of years for NW-migrating silicic volcanism after MacLeod et al. (1976). Solid lines indicate faults and stars indicate Cascade stratocones. Rectangular frame shows outline of Figure II-1b.



Fig. II-1b Simplified geologic map of the western Harney Basin (after Green et al., 1972). Outline shown in figure II-1a. Dense stipple HAOT represents pre-Rattlesnake Tuff high-alumina olivine tholeiite; little stipple HAOT are younger basalts; late-Miocene silicic dome complexes are shown with darkest gray; mid-Miocene domes are identified by a letter M. The postulated source area for the Rattlesnake Tuff (RT) is shown with the dashed circle. SqBu and 209an are andesitic units mentioned in text, and "cont. HAOT" indicates basalt flows contemporaneous with Rattlesnake Tuff. White is mainly alluvium but includes other tuffs and older volcanic rocks.



Geologic and tectonic setting

The Rattlesnake Tuff erupted from the western Harney Basin located on the High Lava Plains of southeastern Oregon at 7.05 Ma (Fig. II-1) (Chapter I, McLeod et al. 1976, Walker, 1979). The Harney Basin was a center of Late Miocene silicic magmatism including numerous silicic dome complexes as well as ash-flow tuffs (Walker, 1970, Walker & Nolf, 1981, Greene, 1972). It is part of a northwestwardyounging silicic magmatic trend, defined by MacLeod et al. (1976), that broadly coincides with the High Lava Plains and the Brothers Fault Zone. Late Miocene -Pliocene magmatism of the High Lava Plains, including Harney Basin, is dominated by thin high-alumina olivine-tholeiite (HAOT) lava flows (Hart et al., 1984).

Eruption of HAOT flows was preceded throughout eastern Oregon by floodbasalt volcanism of the Columbia River Basalt Group to the north and Steens Basalt to the south (both ca. 18-12 Ma) (Hooper & Swanson, 1990, Fuller, 1931, Gunn & Watkins, 1970). The transition zone between the Columbia River and Steens basalt provinces is marked by contemporaneous intermediate calcalkaline volcanic centers striking NW across eastern Oregon (Robyn, 1979, Walker & MacLeod, 1991, Walker & Robinson, 1990, Greene, 1972, MacLean, 1994). Older volcanic rocks in the Blue Mountain Province (Fig. II-1) include the Oligocene-Early Miocene John Day Formation and the Eocene Clarno Formation (Robinson et al., 1990). South of the Brothers Fault Zone, pre-Steens alkaline and calcalkaline volcanic rocks crop out along two large fault scarps and are as old as Middle Tertiary (Langer, 1991, Mathis, 1993).

Analytical methods

Rhyolitic and dacitic Rattlesnake Tuff samples consisted of pumices ranging in size between 40 to 5 cm. All pumices were glassy. LOI (loss on ignition) of rhyolites and most dacites was below 4 weight %. Five of the dacite pumices yielded LOI between 6 and 4 weight %. Major-element analyses were obtained from fused glass discs, with five times as much flux as rock powder, and selected trace elements from pressed powder pellets by X-ray fluorescence (XRF) at Stanford University using a Rigaku instrument. Trace element concentrations were determined by instrumental neutron activation analysis (INAA) at the Radiation Center, Oregon State University using a 1 MW Triga reactor. Analytical uncertainties for XRF trace elements, as based

on replicate analyses of USGS standards G2 and AGV1, are: < 5% for Nb, Zr, Y, Sr, Rb, Ga, Zn, < 10% for Cu, V, Ba, and < 5-15% Ni, Cr, Pb. Uncertainties for INAA trace elements (also based on replicate analyses of in-house standards, CRBIV and SPGa) are: < 5% for Co, Eu, Hf, La, Sc, Sm and Yb, < 5-10% for Ce, Cr, Lu, Ta, Tb, Th, < 5-15% Ba, Cs, Nd, Rb, Zn and < 15% for Ni, U. The range of uncertainties for single elements are based on the concentration range observed in standards used as monitors. Samples analyzed elsewhere are indicated by the footnotes in Table II-1

Basaltic inclusions were analyzed for major and trace elements by neutron activation analysis at OSU. Uncertainties for trace elements are slightly higher due to low weights, but replicate counts of unknowns were performed as a quality check. Especially Zn concentrations were verified, and in sample 60Xcpx, recounted after 15 months with hand calculation due to interference problems. Uncertainties for major and minor elements are on the order of less than 3% for Na and Mn, 5% for Al, Fe, K, and Ti, and 5-10% for Ca and Mg.

Phosphorus concentrations of basaltic inclusions were obtained by microprobe analysis and minerals were analyzed in polished sections and grain mounts on a Cameca SX-50 microprobe at Oregon State University by comparison to known mineral standards.

Eruptive products of the Rattlesnake Tuff

High-silica rhyolites

High-silica rhyolite pumices and shards comprise more than 99% of the Rattlesnake Tuff. At least 5 distinctive high-silica rhyolite compositions are recorded in glassy, white to gray colored pumices (brownish where oxidized); representative analyses are in Table II-1. Banded pumices are common. Glass shards in the matrix of the tuff appear to consist entirely of high-silica rhyolite and vary from white (clear under the microscope) to dark brown. White shards, like white pumices, are concentrated in basal and distal parts of the tuff (Chapter I). White pumices and shards are the most evolved high-silica rhyolites, the gray pumices and shards the least evolved (Table II-1). The distribution of variably colored pumices and glass shards within the deposit is interpreted to reflect the general inverted stratigraphy of a magma chamber zoned from a more differentiated top to deeper less differentiated levels (Chapter I). The chemical zonation of the tuff, as recorded by whole-rock analyses of bulk tuff, is a minimum because of mechanical mixing of high-silica rhyolite compositions during eruption yielding the mixed shard matrix (Davenport, 1971, Chapter I & III).

Analyses of whole-rock pumices indicate that high-silica rhyolites are mostly metaluminous to slightly peralkaline with molar Na+K/Al (agpaitic index) ranging from 0.88 to 1.03 (Table II-1, Fig. II-2). The consequences of possible ion-exchange during post-depositional hydration of glass are discussed in Chapter III. High-silica rhyolite pumices cluster in 5 groups labeled A through E from most to least evolved (Chapter III). Group A to E span a narrow range of major element concentrations (e.g. 3 wt % SiO₂, 1.5 wt.% FeO), but have extreme depletions in Ba, strong depletion in Eu, Sr, LREE (La to Pm), Zr, and moderate depletion in Hf and Zn in group A rhyolites relative to group-E. Rb, Cs, U, Ta, Th, Y, and HREE (Er to Lu) are strongly to moderately enriched in group A relative to group E. Sc seems to have a minimum within group B and C (Chapter III).

Phenocryst contents of pumices range from around 1 weight % for group E to aphyric in group A. The dominant mineral in phyric high-silica rhyolites is a single alkali-feldspar followed by Fe-rich clinopyroxene, quartz, magnetite, fayalite, and trace amounts of biotite (Table II-4). The change in modal mineral content and mineral chemistry towards more evolved high-silica rhyolites is continuous (Chapter III). Trace phases are mainly zircon and apatite which were found in all mineral bearing high-silica rhyolites mainly as inclusions in clinopyroxene and magnetite. Inclusions of pyrrhotite occur in group C to E. Chevkinite (1-3 grains), a LREE-enriched phase, was found in one pumice of group C and D. The abundance of chevkinite is estimated to be one to two magnitudes lower than the most abundant trace phase, zircon.

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Fig. II-2 Silica versus total alkali

All data points are Rattlesnake Tuff samples with divisions according to LeBas et al. (1986). Large filled circles indicate group-E rhyolite, small circles represent other highsilica rhyolite compositions. Squares are dacites to low-silica rhyolites: solid represents dacite with less than 10% non-dacite banding, half-filled represents 10 to 15% banded, and crossed squares for more than 15% non-dacite (see text and Table II-1). For this and all other diagrams, major element data are normalized to 100% anhydrous.



Table II-1 Chemical data

Chemical composition of Rattlesnake Tuff whole rock pumices and related intermediate to mafic lavas, measured at 100% volatile free (no Cl, F, CO₂, or H₂O in samples during analysis). Normalized data used for plotting; n.d.= not detected; -- = not analyzed; italic numbers are INAA values. "%-banded" indicates the textural estimates from handsamples of high-silica rhyolite and basaltic bands of analyzed dacite pumices, number is amount of banding in volume percent: W = white rhyolite pumice bands; G = gray rhyolite pumice bands; M = mafic bands; S = strongly mingled. XRF-data for RT-60A and for #93.2ba are from XRAL company and for HP-33 from Washington State University. Values in parentheses appear anomalous and are not used for plotting. * indicates anomalous high CaO and possibly Sr due to caliche. Pumice identifcation number includes sample locality number and capital letter designates individual pumice clast. Analyses used for average dacite composite are: RT-2A, -6D, -6E, -50A, -50B, -60A, -60B, -60D, -141A, -165C, -210.1A, and -211A. Selected high-silica rhyolites are representative of high-silica pumice groups, HSR-group (Chapter III).

Table II-1:

		high-si	lica rł	yolites			rhyo	lites
No.	RT173H	RT34E	RT165A	RT173B	RT34C	1.0	RT116A	RT116B
HSR-group	A	В	С	D	E	%-banded	S.40G2M	S,30G2M
XRF wt.%								
SiO2	77.43	76.49	76.92	76.22	74.83		72.75	70.06
TiO2	0.11	0.12	0.12	0.14	0.18		0.41	0.66
A12O3	11.85	11.67	11.74	11.91	12.28		12.47	12.76
Fe2O3*	0.87	1.36	1.61	1.92	2.33		3.22	4.58
MnO	0.08	0.09	0.09	0.09	0.09		0.13	0.15
MgO	0.05	0.09	n.d.	0.05	0.07		0.56	1.06
CaO	0.26	0.33	0.35	0.45	0.63		1.33	2.20
Na2O	3.45	3.03	3.69	3.82	3.87		3.08	3.43
K20	5.51	5.9	5.21	4.9	5.36		5.49	4.62
1205	0.01	0.03	0.01	0.02	0.03		0.19	0.29
total	99.62	99.11	99.74	99.52	99.67		99.63	99.81
XRF ppm								
Rb	122	91	80	67	63		73	60
Ba	39	132	380	1230	1914		1320	1450
Sr	2	4	7	12	26		49	83
Zr	175	308	371	433	460		408	421
Nb	39.4	32.0	30.5	27.2	25.4		27.4	25.2
Y	100	96	90	79	75		84	78
Pb	20	19	16	15	15		16	29
Zn	88	105	106	115	117		114	146
Ga	18	18	18	18	18		19	18
V	n.d.	6	2	2	17		14	46
Cu	I	2	2	4	3		12	12
Ni	14	10	10	10	9		12	14
Cr	n.d.	2.4	0.9	n.d.	3		3	10
INAA ppm								
Cs	4.48	3.19	2.80	2.38	2.33		2.85	2.66
U	4.9	3.5	3.1	3.2	2.2		2.3	2.3
Th	9.47	7.71	6.87	6.47	6.12		5.83	5.05
Hf	7.13	9.61	9.86	10.55	11.19		9.99	9.98
Ta	2.16	1.75	1.56	1.34	1.35		1.44	1.37
Sb	1.55	1.33	1.16	1.19	1.34		1.21	1.12
As	4.4	n.d.	5.2	3.9	4.5		n.d.	3.3
Sc	3.93	3.69	3.55	3.76	4.83		8.90	13.78
Co	0.08	0.32	0.08	0.11	0.59		2.65	5.48
La	19.9	38.9	50.6	54.3	49.0		45.9	44.2
Ce	49	90	113	130	106		110	104
Nd	28	4 5	59	68	51		56	56
Sm	9.57	12.89	14.55	14.01	12.26		13.01	12.94
Eu	0.65	1.22	1.47	2.03	2.71		2.58	2.81
Tb	2.22	2.70	2.42	2.12	2.16		2.18	2.15
Yb	10.54	9.66	8.38	7.95	7.51		8.43	7.85
Lu	1.61	1.38	1.28	1.22	1.08		1.27	1.25

			d	acites			
-	PT2A	RT6A	RT6D	RT6E	RT14A	RT14C	RT14D
No.	3W	7W4M	3W5M	1W1M	15G	5	3
%-banded						67 07	66 16
XRF WL. 70	62.67	66.34	68.30	64.76	66.91	07.07	0.5
SiO2	1.25	0.92	0.80	1.11	0.79	0.01	12 58
TiO2	13.29	12.82	12.64	12.94	12.84	5 43	4 76
A1203	7 67	6.24	5.40	7.35	5.63	0.19	0.19
Fe203*	0.25	0.21	0.19	0.25	0.20	0.10	0.65
MnO	1.92	1.64	1.60	1.91	1.32	1.40	6.00*
MgO	4 74	3.09	2.80	3.84	3.40	2.02	3 70
CaO	3 64	3.15	3.31	3.73	4.02	3.80	4.03
Na2O	3.04	4.44	4.32	3.38	3.67	3.90	0.16
K20	0.93	0.46	0.39	0.67	0.37	0.30	08 73
P2O5 total	99.20	99.31	99.75	99.94	99.15	99.52	90.15
VRF nom				10	53	58	53
Rh	50	56	20	1360	1930	1430	1930
Ba	1270	1450	1440	1500	122	107	161*
Sr	162	128	104	155	532	435	634
71	470	514	433	491	24.6	24.5	24.4
Nh	23.5	25.4	25.2	24.5	79	75	74
v	92	88	82	01	40	40	22
Ph	69	74	51	00	169	157	131
70	360	197	155	194	19	19	18.
Ga	19	19	19	20	4.6	60	19
V	79	52	55	00	14	11	17
Cu	15	23	17	17	14	16	12
Ni	18	16	18	10	10	16	6
Cr	20	13	15	10	10		
INAA ppm		0.14	2 34	2.06	2.00	2.05	2.04
Cs	1.98	2.14	2.54	1.5	2.7	2.1	2.0
U	2.7	2.1	4 64	4.02	4.78	4.49	4.77
Th	4.10	4.91	0 82	10.40	11.10	9.89	13.09
Hf	10.21	11.77	1 29	1.23	1.31	1.30	1.29
Ta	1.21	1.37	1.1	1.0	1.217	1.11	1.25
Sb	1.01	1.15	4 3	4.1	3.8	5.0	5.9
As	n.d.	4.5	16.10	24.10	16.52	16.25	15.64
Sc	25.6	20.0	7 40	8.58	5.49	7.44	2.69
Co .	10.34	0.85	7.40			20.5	43.1
1.	413	40.1	40.1	36.2	38.5	39.5	95
Ca	9.2	90	90	84	87	01	48
Le	4.8	50	53	47	43	44	12 2
Sm	13 70	13.14	12.49	12.01	11.37	11./1	3 63
SII	4 66	3.90	3.11	3.85	3.59	3.08	2 13
Th	2 49	2.56	2.10	2.03	2.14	2.12	8 21
Yb	9.07	8.47	7.89	8.04	7.37	1.29	1 16
1.0	1 38	1.31	1.18	1.24	1.09	1.10	1.10

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				dacites			
No.	RT50A	RT50B	RT50C	RT60A	RT60B	RT60D	RT106A
%-banded	2 W	5G	15G	4G	4G	3G	S,20W7M
XRF wt.%							
SiO2	61.35	64.27	66.28	66.02	65.68	65.74	69.53
TiO2	1.40	1.14	0.87	0.89	1.05	0.90	0.66
A12O3	13.67	13.33	12.87	12.72	13.15	12.95	12.90
Fc2O3*	7.74	6.76	5.85	5.63	6.73	6.56	4.23
MnO	0.19	0.17	0.19	0.19	0.19	0.23	0.14
MgO	1.88	1.68	1.79	1.58	1.77	1.37	1.20
CaO	4.46	3.95	4.19	3.56	3.60	3.19	2.85
Na2O	3.50	4.06	3.47	3.99	3.86	4.31	3.78
K20	4.14	3.67	3.71	4.81	3.21	3.38	4.31
P205	0.75	0.56	0.50	0.61	0.52	0.43	0.39
total	99.08	99.59	99.72	100.00	99.76	99.06	99.99
XRF ppm							
Rb	42	50	48	69	49	49	63
Ba	1240	1290	1870	1440	1450	1910	1260
Sr	185	168	163	165	140	143	108
Zr	375	374	495	364	428	621	372
Nb	20.8	22.2	23.7	27	23.4	25.8	25.3
Y	60	64	73	68	73	8 1	78
Pb	15	19	16		24.3	23	15
Zn	135	133	134	130	140	154	125
Ga	18	19	18		19	20	19
V	90	83	55		76	53	103
Cu	23	20	18		19	14	17
Ni	21	13	14		16	14	19
Cr	46	23	13	21	22	15	15
INAA ppm							
Cs	1.64	1.73	1.79	1.89	1.82	1.90	2.50
U	n.d.	2.2	2.3	1.7	1.7	2.9	3.0
Th	4.13	4.12	4.42	4.44	4.32	4.23	4.92
Hf	8.77	8.77	10.57	9.44	9.55	12.60	8.88
Та	1.19	1.21	1.21	1.25	1.18	1.33	1.32
Sb	1.03	0.94	1.10	1.01	0.97	1.16	0.99
As	n.d.	n.d.	5.1	4.3	3.9	4.6	n.d.
Sc	23.96	19.78	17.40	16.60	21.00	22.00	13.06
Co	10.36	8.75	6.33	7.31	9.97	7.24	5.56
La	31.1	34.3	38.8	39.9	40.3	40.1	37.2
Ce	71	80	89	94	89	92	90
Nd	37	35	38	54	50	52	50
Sm	8.89	9.69	10.43	12.40	12.80	13.08	11.94
Eu	3.24	2.96	3.48	3.16	3.44	4.18	2.64
Tb	1.87	1.88	2.05	2.07	2.04	2.13	2.07
Yb	6.50	6.70	7.52	7.17	7.56	8.14	7.90
Lu	0.92	0.97	1.10	1.08	1.10	1.21	1.19

				dacites			
No	RT141A	RT165B	RT165C	RT165F	RT165G	RT2101A	RT211A
%.handed	5G	12G	3W5G	2W8G	15G	2G1M	2G
XRF WL%							
SiO2	68.10	69.80	67.63	69.82	67.96	66.59	65.09
TiO2	0.87	0.58	0.85	0.70	0.82	0.90	1.15
A1203	13.06	12.79	12.90	12.75	12.67	12.87	13.06
F#203*	5.66	5.05	5.72	5.01	5.36	5.87	6.95
MnO	0.18	0.21	0.20	0.18	0.17	0.19	0.21
MaQ	1.31	0.59	1.45	0.94	1.50	1.43	1.97
CaO	2.79	1.98	2.97	2.25	3.02	2.89	3.71
Na2O	3.95	4.45	4.03	4.05	3.82	4.55	4.03
K20	3.61	3.67	3.45	3.76	3.63	3.76	3.19
P205	0.46	0.27	0.51	0.33	0.44	0.46	0.61
total	99.99	99.39	99.71	99.79	99.39	99.51	99.97
XRF ppm							
Rb	56	54	59	61	56	62	55
Ba	1380	2220	1340	1390	1410	1260	1210
Sr	108	99	124	92	112	118	146
Zr	411	638	455	494	401	438	419
Nb	25.3	26.1	26.2	27.3	25.2	26.5	24.4
Y	75	79	78	82	77	80	76
Pb	20	23	34	15	16	4 0	4 5
Zn	136	151	153	140	132	162	162
Ga	18	20	18	19	19	19	19
v	60	20	50	37	56	51	76
Cu	15	7	13	10	15	14	18
Ni	16	9	17	13	15	16	17
Cr	14	1.3	12	10	17	11	16
INAA ppm							
Cs	2.09	2.08	2.34	2.31	2.10	2.46	2.62
U	2.2	2.6	2.5	3.2	2.5	2.8	1.8
Th	4.67	4.86	4.92	5.09	4.77	4.93	4.50
Hf	9.31	12.71	9.76	11.36	9.36	9.87	9.49
Та	1.30	1.32	1.42	1.45	1.27	1.36	1.32
Sb	1.03	1.23	1.02	1.22	0.98	1.10	0.89
As	3.6	4.9	n.d.	4.3	n.d.	4.6	3.1
Sc	16.94	16.73	18.02	16.23	16.00	18.80	22.60
Co	7.60	2.47	6.89	4.93	7.78	6.94	9.76
La	39.9	43.5	37.1	41.4	39.9	36.9	35.7
Ce	96	101	87	95	92	80	86
Nd	53	57	47	52	56	49	48
Sm	12.28	13.00	12.05	12.74	12.33	12.59	12.10
Eu	3.04	3.92	3.47	3.26	2.88	3.16	3.45
Tb	2.12	2.07	2.13	2.20	2.02	2.08	2.07
Yb	7.25	7.87	7.73	8.25	7.66	8.18	7.38
Lu	1.08	1.19	1.18	1.23	1.20	1.24	1.10

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additional dacites andesite contemporan	eous H	HAOT
No. 106Adac RT2101B RT-HM 209an 93.1ba 93.2ba	HP-33	93.5MiSi
monded S S		
VDF with		
STO2 57.35 48.08 48.07	47.77	
TiO2 1.22 0.98 0.92	0.78	
1102 16.13 16.56 16.67	17.49	
A1203 515 3.16 7.28 8.21 11.27 10.91	10.42	10.8
Fe203	0.21	
MnO 4.22 8.81 9.68	10.43	
MgO 6.47 11.66 11.12	11.44	
C_{40}	1.95	2.22
Na20 3.02 5.03 5.04 7.13 0.23 0.25	0.20	n.d.
K20 i.d. i.d. i.d. 0.35 0.11 0.11	0.09	
P205	100 78	
total 100.15 100.55 100.01	100.70	
XRF ppm	2	
Rb 61 66 92 42 1.6 <10	2	1.2.6
Ba 1200 1370 1610 813 156 129	0.5	133
Sr 272 225 320	197	219
Zr 255 51 42	47	
Nb 12.2 1.2 (24)	2.1	
Y 38 24 36	22	
Pb 8.5 2.3	1	
Zn 136 121 236 79 73 98	55	100
Ga 19 17	14	
y 154 262	238	
Cu 42 120	120	
Ni 47 143 205	196	(133)
Cr 22 12 35 66 237 267	233	267
INAA ppm		
Cs 2.37 2.41 2.3 1.87 n.d. n.d.	0.43	0.67
U 2.8 2.9 n.d. 1.7 n.d. n.d.	n.d.	n.d.
Th 5.34 5.58 5.4 3.82 0.20 0.45	0.44	n.d.
Hf 9.22 10.30 11.1 5.89 1.41 1.41	1.13	1.48
Ta 1.34 1.33 1.4 0.73 0.08 0.13	0.067	0.10
Sb 1.39 1.06 1.50 0.21 n.d. 0.24	n.d.	n.d
As 4.7. 4.2 n.d. 1.4 n.d. n.d.	1.1	n.d.
Sc 16.90 10.30 22.3 21.70 47.50 41.70	40.40	42.50
Co 7.02 4.26 8.6 24.50 49.7 48.0	43.40	44.4
La 35.9 46.1 41.2 19.2 3.36 3.24	2.07	3.23
Ce 87 114 100 40 8.0 6.5	5.7	7.8
Nd 48 55 58 21 8.6 n.d.	5.6	4.8
Sm 12.28 13.48 13.25 5.56 2.42 2.15	1.76	2.41
Eu 2.88 2.68 3.87 1.68 0.98 0.87	0.75	0.98
Tb 2.12 2.20 2.32 0.98 0.62 0.51	0.45	0.59
Yb 8 18 8 30 9 11 3.82 2.77 2.56	2.26	2.35
Lu 1.21 1.22 1.38 0.54 0.40 0.42	0.34	0.37

Table II-2 Chemical composition of Rattlesnake Tuff inclusions

All data from INAA except P₂O₅ which was determined by electron microprobe (EMP); number of averaged analyses shown in parentheses. SiO₂ was calculated by difference from 100. Ol & plag inc= chrysolite and plagioclase inclusions; cpx & plag inc.= augite and plagioclase inclusion. 60Xbas represents basalt groundmass material; 60Xplag and -cpx are analyzed plagioclase and clinopyroxene separates making up most of the inclusion. Rb value of 6D.inc2 is anomalously high and is not used for plotting. First number and capital letter of inclusion identification indicates host pumice from which inclusion was extracted. n.d.= not detected, -- = not analyzed. 70% frac= composition at 70% crystal fractionation; major elements model from COMAGMAT (Ariskin, 1992) with starting composition HP-33; trace element model from MXNFRAC (Nielsen, 1990) with starting composition #93.1ba (see text).

	ol & p	lag inc.	cpx	& plag	inc
No.	2A.incl	14D.incl	60Xbas	60Xplag	60Xcpx
INAA wt %					
TIO2	0.72	1.37			
A12O3	22.93	17.36			
Fe2O3*	8.20	11.89	12.66	0.76	9.06
MnO	0.135	0.24			
MgO	3.67	10.06			
CaO	10.87	9.40			
Na2O	2.87	2.70	3.34	3.65	0.45
K20	0.4	0.3			
P2O5. EMP					
total	49.80	53.32			
'SiO2'	50.19	46.68			
INAA DDM					
Cs	0.63	0.65	0.54	0.11	0.08
Rb	28	n.d.	45	3.7	n.d.
Ba	963	670	830	133	n.d.
Sr	n.d.	n.d.	n.d.	514	n.d.
U	n.d.	n.d.	n.d.	n.d.	n.d.
Th	n.d.	1.37	1.00	0.11	n.d.
Hf	1.65	2.87	4.82	0.05	n.d.
Та	0.37	0.45	0.75	0.03	n.d.
Zn	180	122	159	9	23
Sb	n.d.	0.74	0.50	0.15	п.d.
Sc	13.10	24.8	39.1	0.56	154
Co	21.6	52	31.4	0.9	37.7
Cr	96	100	92	11.6	3100
La	12	33	24.7	2.44	1.2
Ce	2.0	57	56	3.5	5.9
Nd	12	42	38	n.d.	8.6
Sm	4.00	10.79	9.70	0.21	3.51
Eu	2.10	3.79	3.79	0.62	1.51
ТЪ	0.82	1.92	1.61	0.06	0.78
Dy	4.3	12.5			
Yb	3.30	6.37	5.34	0.08	2.11
Lu	0.50	0.89	0.78	0.017	0.24

phenocryst-poor inclusions 7							700 5
No.	6D.incl	6D.inc2	106A.incl	106A inc4	116P incl	21012	70%1Fac
INAA wt %					110B.Incl	2101B.ic1	
TiO2	2.08	2.05		2.15	1 40		
A12O3	14.24	13.42		15 60	1.08	2.28	1.80
Fe2O3*	11.56	11.67	9.46	9.79	15.02	13.64	15.92
MnO	0.38	0.44		0.21	10.22	12.56	14.91
MgO	5.03	5.66		3 97	0.28	0.39	
CaO	8.59	7.55		5.07	3.33	4.12	4.68
Na2O	3.68	3.79	3.70	3.09	5.83	6.26	8.31
K2O	0.73	0.78		1.01	3.20	3.57	3.39
P2O5, EMP	1.47(4)	1.09(8)		1.91	2.76	2.67	0.65
total	47.76	46.45		1.22(0)	1.14(6)	1.27(12)	0.29
'SiO2'	52.24	53.55		40.37	43.46	46.76	
				33.43	56.54	53.24	51.25
INAA ppm							
Cs	1.68	2.53	0 99	1.11			
Rb	4 5	(71)	29	37	2.86	1.62	
Ba	917	1056	1142	1051	48	4 5	4.81
Sr	n.d.	n.d.	n d	1031	1087	1182	427
U	n.d.	n.d.	n.d.	n.d.	n.d	n.d.	443
Th	1.94	2.37	1 72	1.0.	n.d	n.d.	
Hf	5.08	6.56	4 80	5.13	2.84	2.50	1.4
Та	0.87	0.94	0.73	0.00	8.20	7.21	3.60
Zn	472	498	145	1.02	1.08	1.06	0.31
Sb	0.81	1.09	0.72	206	457	430	177
Sc	35.9	39 6	35.0	0.92	0.79	0.98	
Co	24.1	22.0	35.0	34.6	33.0	41.3	51
Cr	73	128	20.4	19.7	16.6	19.2	68.6
		120	80	73	4 2	27	1
La	33	35	23	2.0			
Ce	74	80	51	30	37	34	9.89
Nd	48	57	3.2	17	86	77	23.7
Sm	12.70	13 72	0.22	47	49	48	24.6
Eu	4.52	5 24	9.27	11.96	12.8	14.1	6.81
ТЪ	2.54	2 68	1.95	4.45	4.40	5.24	1.93
Dy	16.1	10.6	1.85	2.38	2.11	2.30	
Yb	9.80	10.20	5.50	16.3	12.83	16.41	
Lu	1.49	1 53	5.50	7.48	7.61	7.18	6.54
		1.55	0.76	1.10	1.20	1.08	0.94

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Table II-3 Representative mineral analyses

Sample indicates pumice or inclusion number. Capital letter or number in mineral identification identifies individual grains. Location of analysis of single grains is indicated by: ctr, center of grain; hlf, halfway; rim, near but not directly at rim of grain; pt, point without relative position to grain. n.d.= not detected; -- = not analyzed; An= CaAl₂Si₂O₈, Ab= NaAlSi₃O₈, Or= KAlSi₃O₈, Wo= Ca₂Si₂O₆, En= Mg₂Si₂O₆, Fs= (Fe, Mn)₂Si₂O₆, Fo= Mg₂SiO₄, Fa= Fe₂SiO₄, Te= Mn₂SiO₄; * = resorbed grain. Oligoclase and Fe-augite from "euh. phenos in dacites" used for least-square modelling.

		high-silica	rhyolites	
sample	RT-34E	RT-173B	RT-173B	Rt-34C
feldspar	fsp.D.cntr	fsp.H.hlf		fsp.A.hlf
SiO2	66.06	65.32		65.27
A12O3	19.05	20.05		20.54
FeO	0.20	0.21		0.21
CaO	0.19	0.63		1.13
Na2O	5.78	7.55		8.07
K20	8.13	4.62		3.27
BaO	0.28	1.37		1.51
total	99.69	99.75		100.00
An	0.9	3.17		5.76
Ab	51.4	69.05		74.41
Or	47.6	27.78		19.83
sample	RT-34E	RT-173B	RT-173B	Rt-34C
pyx/oliv	cpx.D.cntr	cpx.I.cntr	ol.I.cntr	cpx.A.hlf
SiO2	49.37	47.84	29.34	48.11
TiO2	0.17	0.26	0.02	0.24
A12O3	0.38	0.37	0.01	0.37
Cr2O3	n.d.	n.d.		n.d.
FeO	21.23	27.75	63.33	27.62
MnO	3.29	2.4	5.44	2.4
ZnO			0.29	
MgO	6.10	2.37	1.51	2.17
CaO	17.82	17.59	0.29	17.75
Na2O	0.66	0.54		0.44
total	99.25	99.34	100.23	99.24
Wo	40.7	41.0		41.0
En	19.4	7.7		4.9
Fs	39.9	51.3		54.1
Fo			3.8	
C T-			06 7	

	-	mafic i	nclusions		euh.	phenos in	dacites
sample	2A.	incl	60x.incl	55incl	RT2A*	RT2101A	RT2101A
feldspar	fspL2.cntr	fspL1.cntr	plagl	fsp.pt7	fspC.rim	fspA.cntr	fspH.hlf
SiO2	48.52	50.82	50.92	51.05	62.33	50.38	54.59
A12O3	32.26	31.05	30.97	31.29	23.73	30.55	27.88
FeO	0.58	0.60	0.62	0.66	0.35	0.65	0.59
CaO	15.57	14.12	13.75	13.74	4.74	13.38	10.59
Na2O	2.53	3.29	3.42	3.43	7.92	3.82	5.17
K20	0.08	0.11	0.14		0.89	0.15	0.25
BaO	n.d.	n.d.			0.45	0.06	0.04
total	99.54	99.99	99.82	100.17	100.41	98.99	99.11
An	76.9	69.9	68.4	68	23.6	65.3	52.3
Ab	22.6	29.4	30.9	32	71.3	33.8	46.2
Or	0.4	0.7	0.8		5.1	0.9	1.48
sample	14Dinc1	2101Ai12	60X.inc1	55inc1	RT2A	RT2101A	RT2101A
pvx/oliv	olDrim	olA.hlf	cpx.2.A	cpx.pt3	cpxC.cntr	cpxA.rim	olF.rim
SiO2	38.7	31.18	50.44	50.35	51.25	49.98	37.54
TiO2	0.02	0.02	1.07	1.13	0.50	1.08	0.01
A12O3	0.02	n.d.	3.34	4.16	1.22	3.62	0.04
Cr2O3			0.47	0.47	n.d.	0.39	
FeO	19.43	56.67	7.73	7.65	17.76	8.27	21.77
MnO	0.32	2.5	0.19	0.2	1.26	0.24	0.38
ZnO	0.01	0.08					0.07
MgO	41.59	9.18	15.53	15.00	10.61	15.34	39.55
CaO	0.28	0.21	19.82	20.62	17.48	20.24	0.24
Na2O			0.44	0.45	0.33	0.41	
total	100.45	99.79	99.08	100.03	100.43	99.57	99.74
Wo			41.8	43	37.2	42	
En			45.5	44	31.4	44	
Fs			12.7	13	31.4	14	
Fo	79	21					76
Fa + Te	21	79					24

Fig. II-3 Mineral compositions

Each point represents one analysis [mol %]. See Table II-3 for representative mineral analyses. Fields indicate mineral compositions of high-silica rhyolites (HSR). Field for HSR-plagioclases (Ab₆₉₋₇₀) is uncertain due to extremely sparse phenocrysts in sample #RT173H (Table III-3). (a) Feldspar in dacites. (b) Feldspar in mafic inclusions. Stars indicate feldspars from chrysolite & plagioclase inclusion #2Ainc1 (stars at lower An are for groundmass crystals), crosses were used for augite & plagioclase inclusions, #55inc1 and #60Xinc1, triangles for phenocryst-poor inclusions, and inverted triangle are single euhedral phenocryst surrounded by mafic selvage. (c) Pyroxene compositions projected onto the pyroxene quadrilateral and olivine compositions projected onto the Mg - Fe+Mn tieline as Fo and Fa+Te components. Vertical bars indicate clinopyroxene and crosses olivine compositions in mafic inclusions; solid circles indicate pyroxene compositions and open squares olivine compositions of single crystals or glomerocrysts consisting of < 5 grains.



Fig. II-4 Textures of mafic inclusions

(a) to (d) are photomicrographs in plane light: (a) augite & plagioclase inclusion with cumulus texture, black between grains is interstitial "magma", field of view is 6.5 mm;
(b) glomeroporphyritic chrysolite & plagioclase inclusion #14Dinc1, plagioclase phenocrysts with hopper texture, field of view is 6.5 mm;
(c) hopper texture of phenocrysts and groundmass of chrysolite & plagioclase inclusion #2Ainc1, field of view is 1.2 mm;
(d) part of phenocryst-poor inclusion #116Binc1 shows mingling texture with host pumice #RT116B, field of view is 6.5 mm.
(e) Backscatter electron image of groundmass of phenocryst-poor inclusion #106Ainc1. Lightest spiny areas are mainly quenched clinopyroxene groundmass crystals, medium gray indicates remaining glass, and black represents vesicles and open areas on grain mount. Bar scale on lower left is 100 microns.

(a)



Fig. II-4, Continued: (b): Top, (c): Bottom





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Fig. II-4, Continued: (d): Top, (e): Bottom



Dacites

Dacite magma in the Rattlesnake Tuff occurs as vitric, commonly black pumices ranging in size from about 2 cm to 30 cm. They are red where oxidized. Uniformly black pumices are rare; most dacite pumices are banded with variable proportions of high-silica rhyolite bands (Table II-1). Dacite pumices often have mafic inclusions ranging from roundish to fine streaks or bands, that are commonly red and therefore contrast well with black dacite. The size and abundance of dacite pumices decreases with distance from the vent similar to the high-silica rhyolite pumices (Chapter I). No dacite pumice was found beyond ca. 100 km. The maximum estimated volume of dacite in the Rattlesnake Tuff is 1%.

When pumices were banded, care was taken to select dacitic parts for analysis through hand-crushing and selecting lapilli size pieces. Some of the non-dacite material could not be removed and its volume % is estimated in Table II-1. Some strongly banded pumices were analyzed as well. Silica contents range from 62 to 73 weight % (normalized, anhydrous; Table II-1). The bulk of analyzed dacite pumices falls between 64 to 68 weight % SiO₂ and when strongly banded pumices are excluded, then 68 weight % seems to be the upper limit (Fig. II-2).

Despite intense sampling efforts, only two small pumices were analyzed that fall in or near the andesite field. The sparseness could be explained by a much smaller size of such pumices, because larger pumices were preferentially collected. Dacites are subalkaline and become mildly alkalic with lower silica content (Fig. II-2). An icelanditic character in the more mafic dacites is indicated by high iron and low magnesium concentrations (Table II-1). Ti, Sc, Co, and Sr are trace elements whose abundances vary linearly well with major element concentrations (Fig. II-9, II-13)

Phenocryst contents in dacites are about 1 to 5%. Heavy-liquid mineral separation on two representative samples yielded 1.2 and 2.4 weight % minerals. Assemblages are strongly bimodal with minerals like those from high-silica rhyolites and those from basaltic magma, even in pumices with little banding (Fig. II-3, Table II-3). The former are in some cases enclosed in a small blob of high-silica rhyolite glass. A selvage consisting of quenched groundmass crystals with interstitial glass is more common around basaltic minerals. Rhyolitic and basaltic phenocrysts are euhedral or indicate signs of disequilibrium through resorption features or reaction rims. Euhedral phenocrysts whose compositions cannot be matched either with high-silica rhyolite or basaltic minerals are likely to have crystallized from dacite magma. Such phenocrysts

consist of augite (En₂₅₋₃₅Wo₃₈) (Fig. II-3, Table II-3). Anhedral plagioclase (An₃₀₋₅₀) may have also crystallized from dacite magma (Fig. II-3, Table II-3).

Basaltic inclusions

Basaltic magma makes up less than 0.1% of the eruption products of the Rattlesnake Tuff. Mafic magma inclusions occur dominantly in dacite (c.f. Davenport, 1971, p. 88) or dacite-rhyolite banded pumices. Only three very small, high-Mg olivines were found in one completely rhyolitic pumice.

Basaltic material occurs as single phenocrysts, with or without selvages of quenched groundmass crystals with interstitial glass, as round or ellipsoidal bodies, or as fine streaks with typical length-to-thickness ratios of 10 to 20. The inclusions typically range from 3 cm to less than a mm long. All of the basaltic inclusions are interpreted to be cognate based on a combination of: (1) elongate or quenched inclusion shapes indicating they were liquid; (2) quench textures of phenocrysts or in the inclusion groundmass, and (3) uniformity of textures, modes, and chemical composition of olivine and plagioclase within types (Fig. II-4).

Among the inclusions that are > 0.3 mm long, three types are distinguished: phenocryst-poor inclusions; inclusions which contain glomerocrysts of olivine and plagioclase; and granular (or cumulus) inclusions which consist of $\geq 90\%$ crystals. Phenocryst-poor inclusions indicate the strongest mingling textures with the host pumice. The quenched nature of the groundmass is clearly revealed only with help of backscatter electron images (Fig. II-4). Plagioclase phenocrysts are often tabular. Inclusions with glomeroporphyritic chrysolite and plagioclase are always roundish and commonly have stronger quenching textures of groundmass and plagioclase phenocrysts suggesting supercooling on the order of 100 to 200 °C (Fig. II-4) (Lofgren, 1980). Some of glomeroporphyritic inclusions have textures transitional to the granular inclusions with crystal content reaching 50% or more. The granular (or cumulus) inclusions consist of either Fe-hortonolite and plagioclase or augite and plagioclase assemblages with some Fe-Ti-oxides. All granular, augite and plagioclase inclusions have plagioclase phenocrysts poikilitically enclosed by augites with some interstitial groundmass left (Fig. II-4). The largest augite and plagioclase inclusion is about 15 cm in diameter and is now a vug lined containing dozens of cm-size augite and plagioclase inside a large dacite pumice. The largest single clinopyroxene is 8 mm across.

Plagioclase compositions in all inclusions range between An₇₅ to An₅₀, typically with zonation of less than 5 % An (Fig. II-3). The most calcic feldspars are phenocrysts in the glomerocrystic and granular augite and plagioclase inclusions. Mgrich olivines ranges are Fo₈₀₋₇₈ and Fe-rich olivines are Fo₂₃₋₂₂ (Fig. II-3c). Within single host pumices, the olivine in chrysolite and plagioclase inclusions ranges from fresh olivine to completely altered to iddingsite. Single euhedral chrysolite crystals surrounded by dacite are identical to olivines from glomeroporphyritic chrysolite and plagioclase inclusions. Poikilitic augites have restricted compositions around $Wo_{42}En_{44}Fs_{14}$. and contain ca. 0.4 weight % Cr₂O₅ (Table II-3, Fig. II-3c). Compositionally identical augites and augites with slightly more Fs and less Wo are also found as single euhedral phenocrysts in dacite pumices. Such phenocrysts commonly have a selvage or contain inclusions consisting of quenched groundmass crystals with interstitial glass (Table II-3, Fig. II-3c).

Two glomerocrystic inclusions, six aphyric (phenocryst-poor-type) inclusions, and all the components of one cumulus augite and plagioclase inclusion where analyzed for major and trace elements (Table II-2). One chrysolite and plagioclase inclusion shows chemical evidence for accumulation of feldspar (high Al_2O_3 and low $Fe_2O_3^*$ and Sc in inclusion 2Aincl) and the other possibly for accumulation of olivine (high Co combined with low Sc in 14Dinc1).

Discussion

Rhyolite genesis

High-silica rhyolites of the Rattlesnake Tuff are closely pre-and postdated by other Late Miocene rhyolites erupted as ash-flow tuffs or as lavas (some Domes) from the Harney Basin (Fig. II-1b) (Green, 1973, MacLeod et al., 1976 Walker, 1979, MacLean, 1994). The whole suite of rhyolites from the Harney Basin comprise low- to high-silica, metaluminous to mildly peralkaline rhyolites. Mineralogically and with respect to trace-element composition they can all be classified as subalkaline, A-type volcanics using the scheme of Eby (1990). High-silica rhyolites (> 75 %SiO2) are distinct in their low MgO (< 0.2 wt%) and CaO (< 0.8 wt%) contents combined with molar (Na+K)/Al of greater than 0.9. Low-silica rhyolites are more variable in major element composition than the high-silica rhyolites.

The variability of the high-silica rhyolites is subject of chapter III. The discussion here concerns only the least evolved high-silica rhyolites, group E rhyolites. Group E rhyolites are considered least evolved because they have the highest Fe, Ti, Sr, lowest Si, Rb, Th, and the smallest negative Eu anomaly. (Chapter III). We here consider their possible origin by fractionation, or by partial melting of a suitable protolith, or by some combination of processes.

Fractionation models

Three parental compositions were considered for crystal fractionation models to derive group E rhyolites: regional rhyolites, Rattlesnake Tuff dacites, and HAOT. Group E rhyolites cannot be derived from regional rhyolites based on comparison of Rb, Cs, Th, U, and Ba. On variation diagrams of silica versus Ba, Rb, and Th (and implicitly Cs and U), group E rhyolites have lower Th and Rb and higher Ba concentrations than Harney Basin rhyolites. High Ba (1800 to 2000 ppm) cannot be ascribed to accumulated phenocrysts; heavy-liquid mineral separations on one representative sample (#RT173C) yielded only 0.5 weight percent anorthoclase, which contributes 64 out of the 1834 ppm Ba in the sample (Chapter III). No equilibrium phase assemblage exists for rhyolite in which Rb is highly compatible and Ba incompatible. Indeed, with alkali feldspar as the dominant phase, the reverse is true and group E rhyolite is potentially a parental composition for regional rhyolites, rather than the reverse.

The dacites also prove to be unsatisfactory parental compositions based on: the misfit of fractionating proportions between major and trace element models; incompatible trace element variations that are internally inconsistent with fractionation; and because dacite itself appears to be a derivative of mixing between basalt and group E rhyolite. Most crystals in dacite belong either to rhyolite or basalt. Least-squares mass balance calculations for major elements (after Morris, 1984), using the phenocrystic dacite assemblage oligoclase, Fe-augite (see table II-3), and titanomagnetite (see #RT173C table II-3), yield estimates of 32 to 40 percent fractionation from dacite to yield group E rhyolite with sum of the residuals squared of 0.21 to 0.25. Including two additional phases, orthopyroxene (pigeonite) and ilmenite, increases the amount of fractionation to 35 to 42 percent and improves the fit to 0.07 to

0.09. In each case at nearly identical fits, more fractionation was achieved for the slightly more mafic average dacite composition than for the dacite #RT141A (Table II-1). P2O5 was not considered.

The amount of Rayleigh fractionation consistent, within analytical error, with the variation in Rb, Th, U, La and Ce from dacite (avg. dacite and #RT141A dacite) to group E rhyolite is at most 30 percent, using a fairly high estimate of 0.3 for the bulk partition coefficients. For Nb and Ta which are also highly incompatible in slightly peralkaline rhyolites (Chapter III, Mahood, 1981), bulk partition coefficients would have to be 0.8 for 30% fractionation, which is unrealistically high. Smaller degrees of fractionation can satisfy the trace elements Ta and Nb, but aggravate the misfit for other trace elements and the disagreement between major and trace elements.

Major element modeling was done in two steps to derive group E rhyolite from HAOT. The first step included the fractionation from a primitive HAOT (#93.1ba) to a more evolved mafic composition using a forward model program (Nielsen, 1990) with perfect fractionation and oxygen fugacities at FMQ. Modeling the second step, evolved basalt (70% fractionate of high-Mg basalt #93.1ba) to group E rhyolite, mass balance calculations by least-square methods (after Morris, 1984) yielded 79 percent fractionation of an assemblage of 30% augite, 31% labradorite, and 13% titanomagnetite with an unacceptably high sum of squared residuals of 3.1. With additional pigeonite and ilmenite, fractionation amount and fit are identical. Crystal fractionation models for trace elements suggest fractionation amounts of \pm 95% using Rb, Th, Nb, Ta, La, and Ce (D= 0.1) to achieve concentrations observed in primitive HAOT's with starting composition group E rhyolite. In conclusion, arguments against crystal fractionation models from HAOT include the misfit of the least square models and the lack of evidence for fairly primitive intermediates that would lie along the crystallization path.

Partial melting models

Major element compositions of dehydration melts of natural rocks at various pressure and temperature conditions share similarities with group-E rhyolites (Fig. II-6). Protoliths for dehydration melting experiments include greenschists and a mafic hornfels (Beard & Lofgren, 1991), a tonalitic gneiss (Skjerlie & Johnston, 1993), and high-grade metapelites (Vielzeuf & Holloway, 1988, Patino-Douce & Johnston, 1991). the first

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Fig. II-5 Scatter diagrams of Rattlesnake Tuff and other Harney Basin rhyolites

Large and small solid circles indicate Rattlesnake Tuff samples, large for group-E rhyolites and small for representative more evolved high-Si rhyolites. Dotted lines indicate fractionation trend within Rattlesnake Tuff high-silica rhyolites. Open circles show other Harney Basin tuffs and open squares silicic lava flows from Late Miocene dome complexes. Sources: Harney Basin Tuffs, Streck, unpub. data; silicic domes: MacLean, 1994; Grunder, unpub. data.



Fig. II-6 Dehydration melts over average group E rhyolite

Comparison of Rattlesnake Tuff average group-E rhyolite to rhyolite dehydration melts. Values above 1 indicate experimental melt was richer in component and below 1 indicate experimental melt was poorer. Values of greater than 10 or less than 0.1 not displayed. a) melts from tonalitic gneiss of Skjerlie & Johnston, 1993, b) melts from meta-basalts of Beard & Lofgren, 1991 c) melts from meta-pelites of Vielzeuf & Holloway, 1988, d) melts from meta-pelite of Patino-Douce & Johnston, 1991. In a) to d), the melt with the greatest similarity to group E rhyolite is shown with the solid circle.



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Fig. II-7 La_N/Yb_N versus Ba/Rb of group E rhyolites and potential source rocks

Open squares represents group-E Rattlesnake Tuff rhyolites, solid dots show representative analyses of HAOT and Steens-basalt (Table II-1, Gunn & Watkins, 1970, Helmke & Haskin, 1973); primitive HAOT's always have LaN/YbN <2, but can have higher Ba/Rb as shown. Solid triangles represent mid-Miocene Harney Basin dacite domes with likely plutonic counterparts, in the vicinity of the Rattlesnake Tuff source area (MacLean, 1994); solid diamonds represent mid-Miocene eastern Oregon andesites (Strawberry volcanics, Robyn, 1979). Steens Mountain basalt and Strawberry andesites, both 15 Ma old, might also make up some mid-crustal basement rocks as amphibolitized metamorphic rocks . Average I-type granitoid taken from Chappell & White, 1992. Crosses indicate pelites (NASC and Meta-NASC, Gromet et al., 1984) and argillites (Hill et al., 1981). Solid crosses used for greywackes (Hill et al., 1981, Taylor & McLennan, 1984, p. 84, #M285). I-type granitoid and sedimentary rocks represent components likely to be present in and between the accreted arc terranes postulated to underlie the region (Brooks & Vallier, 1978; Dickinson & Thayer, 1978). Star indicates proposed average lower crustal composition (Taylor & McLennan, 1984, p. 92). Vectors approximate proposed elemental change during transition from amphibolite- to granulite-facies metamorphism (Fowler, 1986, Taylor & McLennan, 1984, p. 78) and when residual garnet is left in the melting region (Leake, 1990).



Despite the large silica variations of the protolith (50 to 68 % SiO2), silicic melts (>71% SiO2) were generated from each protolith at melt percentages ranging from 8 to 60%. Compared to average group-E rhyolite, alumina is only slightly too high in all of the experimental melts, whereas in hydrous experiments, alumina is much greater. Magnesium in all and calcium concentrations in about half of the experimental melts are much higher and sodium and potassium commonly lower than observed in average group-E rhyolite. Iron, titanium and manganese do not deviate consistently. Phosphorus was not compared, due to analytical uncertainties at very low concentrations.

The best compositional matches are for melt from basaltic greenschist at 6.9 kb and 850 °C (protolith #478 of Beard & Lofgren, 1991) and for melt derived from tonalitic gneiss at 6 kb and 975°C (Skjerlie & Johnston, 1993). The elemental deviation of the closest metapelite melt was not considerably worse (Fig. II-6). It is therefore conceivable that dehydration melts could produce group-E rhyolite composition except possibly the low magnesium concentration. A suitable protolith might be of intermediate composition containing 3% sodium and 1-2 % potassium at modest magnesium and calcium concentrations.

Trace element data do not exist for dehydration melt experiments, but some simple trace element tests can be constructed to examine the plausibility of dehydration melting for the origin of group E rhyolite. Most important is the high concentration of Ba in group E rhyolites at relatively low Rb concentrations, which indicates that these rhyolites have not undergone much fractionation in modal proportions of phases similar to the ones observed. The bulk partition coefficient for group-E rhyolite is 5.7 for Ba and only 0.1 for Rb, based on mineral separate data (Chapter III). Parent rhyolite would have to have higher Ba at even lower Rb concentration than the group-E rhyolites, which is unrealistic. It also follows that any suitable protolith must have had a Ba/Rb ratio equal to or greater than that of group-E rhyolites because Ba is equally or less incompatible than Rb during melting (Fig. II-7).

Similarly, La_N/Yb_N ratios, reflecting the steepness of the rare earth element pattern, help restrict potential protoliths. Partial melting produces melts with La_N/Yb_N ratios as great or greater than the parent, provided residual LREE-enriched trace phases are not governing the REE systematics. This condition is met in most cases of crustal melting, including those of the experimental dehydration melts.

Among geologically likely basement constituents, only Steens Mountain basalt, HAOT, and some greywackes, are suitable source rocks as long as garnet is not a major residual phase during melting (Fig. II-7; Leake, 1990). Metamorphic conditions ちょうしんかち からて 二十二日 「日日日日 あるのの や や

could potentially transform some unsuitable basement rocks by increasing their Ba/Rb and decreasing their La/Yb ratios during granulite facies metamorphism (Fowler, 1986).

In summary, we favor the generation of the least evolved Rattlesnake Tuff highsilica rhyolites (group-E rhyolites) through dehydration melting of ca. 10% of an amphibolitic rock representative of fractionated basalts (low MgO, CaO, TiO2) or 40 to 50% melt of a meta-greywacke, with a pelitic component to increase the bulk K₂O content. This model is also in accordance with provenance and tectonic models for southeastern Oregon, which includes, terrane amalgamation since early Cretaceous, and with Tertiary magmatic evolution (Dickinson & Thayer, 1978, Brooks and Vallier, 1978, Wells & Heller, 1988).

Mafic root zone

Contemporaneous and regional basalts

Basalts younger than about 11 Ma in the northwestern Great Basin, including the Harney Basin, are high alumina olivine tholeiites (HAOT) (e.g. Hart, 1984). Individual lava flows are commonly 5 to 10 m thick. Their petrographic and geochemical characteristics are similar over the whole distribution area from northeastern California, across southeastern Oregon to the tristate border of Oregon, Idaho, and Nevada (Carlson & Hart, 1981, McKee et al., 1983, Hart, 1984, Grove et al., 1988, Draper, 1991, Bailey & Conrey, 1992). Phenocrysts, if any, consist of olivine with Fo₆₈₋₈₈ and sometimes plagioclase ranging from An₆₀ to An₈₄ (Hart, 1984, Draper, 1991, Grove et al., 1988). Augite is the only pyroxene and is restricted to the groundmass (Hart, 1984). The geochemical characteristics are high MgO/FeO* at MgO concentrations commonly between 8 to 10 weight %, low K₂O (< 0.3 weight %), and low incompatible-element concentrations, which make them similar to mid-oceanridge basalts and back-arc basin basalts (Hart et al., 1984, Goles, 1986, Draper, 1991, Bailey & Conrey, 1992).

HAOT lava flows with the same characteristics erupted before and after the eruption of the Rattlesnake Tuff in the Harney Basin and range in age from about 8 to 2.5 Ma (Fig. II-1b) (Parker, 1971, Walker, 1979). That HAOT magma erupted also throughout the lifecycle of the Rattlesnake Tuff magma chamber is documented 40 km

southwest from the center of the tuff's source area (Fig. II-1b). There, basalts, with chemical signatures typical for the most primitive HAOT's from Oregon (Table II-1) (Goles, 1986, Draper, 1991, Bailey & Conrey, 1992), occur beneath, above, and are contemporaneous with Rattlesnake Tuff. The Rattlesnake Tuff (7.05 Ma) directly overlies the pahoehoe surface of a basalt (HP-33) and is in turn overlain by basalt (93.1ba), which is capped by 6.78 Ma old Buckaroo Lake Tuff (Grunder & Deino, unpub. data). Also at the same location, there are several basalt intrusions in the Rattlesnake Tuff. One of these sill-like bodies is truncated by faults which are themselves filled with rheomorphic Rattlesnake Tuff indicating that primitive HAOT magma erupted even during the emplacement of the Rattlesnake Tuff.

Mafic inclusion - HAOT connection

Some of the Rattlesnake Tuff mafic inclusions are compositionally and mineralogically like HAOT basalt flows and others are distinctly different (Table II-2, Fig. II-4). All phenocryst-poor inclusions represent more evolved basaltic andesite compositions, with lower magnesium, calcium, aluminum and higher sodium and potassium concentrations than HAOT (Table II-2). The phenocryst-poor inclusions follow a mild iron-enrichment trend (Fig. II-8). Minor and trace elements indicate a tholeiitic differentiation trend by enrichment in TiO₂, MnO, P₂O₅, and Zn (Table II-2) (Carmichael et al. 1974, Naslund, 1989). Evolutionary trends of dacites also point back to a *fractionated* tholeiitic liquid as mixing member or fractionation parent with precisely the characteristics of the observed phenocryst-poor inclusions and not mixing with a primitive HAOT (Fig. II-8 and II-9).

Inclusions with chrysolite and plagioclase glomerocrysts are mineralogically identical to phenocrysts from primitive HAOT (Table II-3) (Hart, 1984, Grove, 1988, Draper, 1991) and their major element compositions are similar to HAOT's. Their chemical character is likely to be partially influenced by the high olivine and plagioclase phenocryst content of the two inclusions analyzed. Inclusion 14Dinc1 has trace element concentrations intermediate between HAOT and the phenocryst-poor inclusions providing further evidence for differentiation of the basaltic andesites from HAOT and that HAOT magmas interacted with magmas of the Rattlesnake Tuff chamber (Fig. II-9 and II-14).

Fig. II-8 AFM-diagram

A, Na2O + K2O; F, FeO*; M, MgO, all in weight %. Solid line separates tholeiitic from calcalkaline field after Irvine & Baragar (1971). Solid diamonds represent Rattlesnake Tuff mafic inclusions, except inclusion #2A.inc1 not plotted as in all following figures; solid squares indicate representative Rattlesnake Tuff dacites and solid circle for average group-E rhyolite. Open diamonds show contemporaneous HAOT's and cross shows 6.71+/- 0.08 Ma (whole rock K-Ar-age, unpub. data) andesite flow #209an which is also shown in figure II-1b and II-15 and listed in table II-1. Mix. HAOT-HSR: mixing line between primitive HAOT and Rattlesnake Tuff high-silica rhyolites. None of the phenocryst-poor inclusion falls on the mixing line rhyolite-HAOT or is compositional similar to HAOT, except chrysolite & plagioclase inclusion #14Dinc1.



Fig. II-9 SiO2 versus. P2O5 and TiO2 for Rattlesnake Tuff samples Symbols as in figure II-2, Contemporaneous HAOT basalts represented by ruled field. All inclusions are phenocryst-poor inclusions except chrysolite & plagioclase #14Dinc1.


The large poikilitic augites of the clinopyroxene & plagioclase inclusions are similar to groundmass augites from HAOT's (Table II-3) (c.f. Grove et al., 1988). Only granular, Fe-hortonolite & plagioclase inclusions are mineralogically and texturally different from HAOT's. Olivines are Fe-rich (Fo₂₁, Table II-3) and are similar to olivines described from a basaltic andesite cone (Squaw Butte, MacLean, 1994) about 10 km from the vent of the Rattlesnake Tuff (Fig. II-1b). Such Fe-rich olivines are also consistent with extreme differentiation along a tholeiitic trend.

Granular inclusion mineral compositions can be matched with euhedral phenocrysts which are surrounded by selvages or contain melt inclusions (Table II-3). Selvages and melt inclusions are texturally identical and consist of quenched groundmass crystals and glass. The compositional match between minerals of the granular inclusions and single euhedral phenocrysts suggests that they represent crystal accumulation or growth along walls and mineral precipitation in magma, respectively. Only the ferro-hortonlites were not found as individual phenocrysts.

Modeling

Major and trace element modeling was done to test whether the basaltic andesite composition represented by the phenocryst-poor inclusions could be generated by fractionation from a primitive HAOT magma as recorded by the contemporaneous HAOT's. Some results are displayed in figure II-10 and II-11.

Crystal fractionation using the forward model COMAGMAT (Ariskin, 1992) were considered first to model major element compositions of phenocryst-poor inclusions and to compare observed with modeled mineral assemblages. Composition HP-33 was used as parent composition at three pressure 0.1, 0.4, and 0.7 GPa, and one run included 0.15 weight % water. The oxygen fugacities were set to FMQ for all runs. The best matches were runs with 0.15 weight % water at pressures \leq 0.4 GPa (Fig. II-10); a pressure of around 0.4 GPa is thought to be most likely if basaltic magmas fractionated underneath a shallow (ca. 0.1-0.2 GPa) rhyolitic cap of the Rattlesnake Tuff chamber. Fractional crystallization of ca. 70 % reproduces most of the observed major element concentrations in the phenocryst-poor inclusions (Fig. II-10, Table II-2). Exceptions are K concentrations of more than 0.8 weight % and high P and slightly lower FeO* concentrations observed in all phenocryst-poor inclusions. Observed and modeled major phases agree well (Table II-3, Fig. II-10). 70% crystallization is also close to the saturation of titanomagnetite and high values of Zn, which is strongly compatible in titanomagnetite, in the inclusions suggest titanomagnetite was not yet, or barely fractionated. Orthopyroxene, which the model predicts, was found in only one sample, however the observed composition (Fig. II-3c) closely corresponds to predicted compositions En75-67Fs19-27Wo₆.

To model trace elements, the program MIXNFRAC and BLF (Nielsen, 1990) was used. Both programs produce similar major element results as COMAGMAT but include experimental and empirical partition coefficient data for simultaneously modeling trace elements. Models where fractionation occurs throughout the magma chamber, known as homogenous fractionation (Nielsen, 1990), was modeled with MIXNFRAC and fractionation along a boundary layer at the chamber walls with BLF. Modeling is limited to 1 atm anhydrous. Composition #93.1ba was used instead of HP-33 because although virtually identical in major elements, it is slightly more enriched in incompatible trace elements therefore reducing slightly the compositional difference between inclusions and primitive HAOT's (Table II-1). Other modeling conditions were: (i) oxygen fugacity at FMQ buffer, (ii) perfect fractional crystallization.

At 70 % homogeneous or boundary layer crystal fractionation, modeled incompatible trace element concentrations are mainly two to three times lower than observed concentrations (Table II-1, Fig. II-11). In case of boundary layer fractionation, varying the degree of fractionation in the boundary layer from 50 to 90 to 99% before backmixing the remaining boundary layer liquid into the main magma of the chamber has a slight effect on the slope of the pattern but does not change the overall enrichment at 70% magma chamber fractionation (cumulative solidification percent of boundary layers relative to size of chamber). Therefore fractionation alone seems incapable to yield element concentrations of the basaltic andesite inclusions. However, overestimating the partition coefficients in the boundary layer at high fractionation amounts of greater than 90 % could cause the modeled concentrations to be lower. Trace element concentrations in the boundary layer are much higher using bulk partition coefficients for an assemblage of 50% plagioclase, 30% augite, and 20% olivine compared to the concentrations obtained for the boundary layers with program BLF (Fig. II-11c & d).

Fig. II-10 Results using forward model COMAGMAT-3.0 (Ariskin, 1992)

Starting composition is HP33 with oxygen buffer at FMQ and conditions as displayed on figure. Dotted horizontal lines were superimposed and indicate results at 70% crystallization which reproduced best observed major element variations in phenocrysts-poor inclusions (Table II-2).



Fig. II-11 Trace element systematics of mafic inclusions

a) Normalization of phenocryst-poor mafic inclusions over primitive, contemporaneous HAOT (#93.1ba, Table II-1). Range of other contemporaneous HAOT's shown by field.

b) Trace element modeling. 70%X-frac: relative concentrations at 70% crystallization using $Eu^{2+}/Eu^{3+} = 0.5$, dotted lines with $Eu^{2+}/Eu^{3+} = 1$ (see text and table II-2). Other relative concentrations represent recharge models (O'Hara, 1977) starting with composition 70%X-frac: solidified mass fraction per cycle 0.5; mass fraction erupted per cycle 0; mass of recharged magma (HAOT #93.1ba) 0.5. Empirical partition coefficients used for plagioclase and olivine from inclusion 60X; other coefficients for plagioclase: K_{Nd} =0.01; for clinopyroxene: K_{Rb} , Ta, Th =0.01, K_{Hf} =0.3; for olivine: K_{Tb} =0.013, K_{Yb} =0.023, K_{Lu} =0.026, K_{Zn} =1, K_{Sc} =2, K_{Co} =20, K_{Cr} =6, all other $K_{element}$ =0.01. 2nd-OI & Plag: indicates composition at the end of the second cycle with olivine and plagioclase fractionation; 14th-Plag & OI & Cpx: composition at the end of 14th cycle with fractionation assemblage 50% plagioclase, 30% clinopyroxene, and 20% olivine. 14th-Plag & OI & Cpx & "Opx": addition of opx to previous mineral assemblage was approximated by increasing coefficients for clinopyroxene of K_{Yb} from 0.4 to 0.5 and of K_{Lu} from 0.31 to 0.52; one other change during this run included the reduction of K_{Zn} for olivine from 1 to 0.5

c) Dotted lines represent average of least evolved (grp-E) and average of most evolved (grp-A) Rattlesnake Tuff high-silica rhyolites; 10%mix.comp is calculated hypothetical composition to derive average of phenocryst-poor inclusion by mixing of 10% contaminant (10%mix.comp.) and 90% of composition 70%X-frac (see Fig. II-11b). 90%- and 99%frac-boundary is composition of boundary layer liquid after 90% and 99% crystal fractionation, respectively, using mineral assemblage 50% plagioclase, 30% clinopyroxene, 20% olivine and partition coefficients as in b).

d) Boundary layer fractionation results using program BLF (see text): 70%BLF@xx% = magma chamber composition after 70% crystal fractionation with 50 to 99% crystal fractionation in boundary layer; 1stBLxx% = composition of first boundary layer liquid after 50 to 99% fractionation.

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Other means to increase incompatible trace element budgets while buffering major elements and slightly reducing compatible trace elements relative to starting composition #93.1ba include recharge models (O'Hara, 1977). Recharge models incorporate cycles in which each begins with an initial liquid which fractionates. After a selected amount of fractionation, the initial liquid composition is recharged into the same chamber where it mixes with the fractionate. One cycle is complete after this mixing event and continues with the next cycle during which the mixed liquid starts to fractionate again. Some results are in figure II-11b. Recharge was modeled with the composition derived from the MIXNFRAC model at 70% fractionation as starting (initial) composition and composition #93.1ba as the liquid which is recharged. 50% was chosen as fractionation amount of each cycle. Empirical partition coefficients for plagioclase and clinopyroxene were derived from the components of one granular inclusion and compare well with published ranges for partition coefficients (Table II-2) (summary of Henderson, 1984, p. 91; Gallahan & Nielsen, 1992). Recharge models with 14 or more cycles come close to the observed concentrations of trace elements in phenocryst-poor inclusions with some exceptions (Fig. II-11). Observed concentrations of Rb, Ta, La, Ce, and, in some cases, Th in inclusions are still higher than models can produce with a parent compositionally as primitive as the contemporaneous HAOT's. Eu concentrations in the inclusions cannot be reproduced using plagioclases with positive Eu-anomalies as observed in plagioclases from inclusion #60X (Table II-2), despite using different Eu2+/Eu3+ ratios for MIXNFRAC (Fig. II-11).

Another possible explanation for the enrichment of incompatible trace elements in the inclusion is contamination by trace element-rich assimilants. Rattlesnake Tuff rhyolites are unsuitable contaminants because they contain either lower concentration of incompatible elements or, if higher, are not high enough because contamination would only yield the desired effect at percentages greater than ca. 20 % which in turn would disturb the major element composition of the inclusions. In order not to disturb the major element composition of the inclusions by felsic assimilants, the required contamination has to occur at low percentages and therefore the contaminant has to have very high trace element concentrations. A hypothetical contaminant to yield compositions of the inclusion through 1 to 9 mixing proportions with 70%-fractionate of #93.1ba (Fig. II-11b) is shown in figure II-11c. Interestingly, the concentrations of the most incompatible elements, Rb, Ta, Th, Ba, La, and Ce, of the hypothetical contaminant are close to the composition of the 99% boundary layer fractionate using fractionation assemblage plagioclase, augite, and olivine. This implies that a boundary layer may potentially operate as source of liquids with high trace elements concentrations boosting the incompatible elements of the main magma during backmixing as would occur during contamination by trace element rich crustal components. Despite the lack of a single model to explain trace element concentrations, the composition of the phenocryst-poor basaltic andesite inclusions is likely a product of complex interplays of recharge, homogenous and boundary layer fractionation and does not require much crustal assimilation

Dacites: mafic-silicic interface

Mixing - mingling and hybridization

There is abundant evidence that dacite in the Rattlesnake Tuff was formed by mixing between rhyolite and underlying differentiated mafic material. Evidence includes: the strong bimodality of phenocrysts into those like phenocrysts in rhyolites and those like phenocrysts in basalt; the colinearity of differentiated basalt-dacite-rhyolite on discrimination diagrams (Fig. II-2, II-8, II-9, II-14); and the occurrence of dacite at the rhyolite basalt interface. The latter is inferred from: (1) dominance of dacite pumice proximal to the source suggests dacites came from deeper in the chamber; and (2) abundant quenched mafic inclusions and mafic phenocrysts inside dacite pumices indicate basaltic magma was underneath.

To test the mixing hypothesis, we calculated the amount of mixing necessary for the generation of two representative dacite pumices and a dacite composite (Fig II-12). Only pumices with the least amount of banding were used for the composite (Table II-1). One of the mixing partners is a composite of the phenocryst-poor inclusions. The rhyolitic mixing member is difficult to approximate because of the variability of Rattlesnake Tuff high-silica rhyolites. Assuming the least fractionated rhyolites are deepest in the chamber, we chose average of group-E rhyolites.

With respect to major elements, the required mixing proportions are 40 to 65 % for dacite with 63 and 68% silica, respectively with an average of 55% for the dacite composite. Proportions are consistent with most trace elements. Na, Cs, Sm, Tb, Yb, and Lu are not shown in figure II-12, because their concentrations are nearly identical in the two endmembers, so mixing proportions cannot be determined. Sr and Pb data were not available for the inclusion. Pb, for example seems to peak within the dacites

and may not fit the hybridization model. Hybridization of a mixture of 50% rhyolite and 50% mafic magma seems a valid model to form dacite.

Despite compelling evidence for mingling through banded textures, the mixing processes persisted long enough to produce a true hybrid dacite. Dacite pumice is texturally distinct, with a dull luster and a scoriaceous appearance, and not merely an intimate banding of rhyolite and basalt. Limited microprobe analyses on several dacite pumices yielded a silica range of 65 to 67 weight percent. It is difficult to evaluate how much of the compositional scatter observed within the dacite pumices is due to gradients within the dacite magma and how much can be ascribed to late stage mingling phenomena. We estimate a maximum range of 63-68 weight % for hybrid dacite mainly based on textural evidence. The lower limit of the estimate is set by the most mafic pumices; the most mafic pumice (RT-50A) shows no evidence for banding of mafic magmas (Table II-1). Pumices with little banding reach silica concentrations as great as 68 weight %. All pumices above 68 weight % SiO2 have stronger banding suggesting that this is the upper silica boundary of dacite magma (Table II-1). The contribution of a "mafic component" through non-dacite mafic minerals is negligible because mineral contents of dacite pumices is on the order of 1-5 %. For example, in sample RT-50A, as in other dacite pumices, high-Cr augites were found sporadically but a Cr content of 46 ppm limits the Cr-augite content to 1.5 volume % of the bulk pumice assuming an extreme case where all chromium came from the augites (Tables II-2 and II-3). The best evidence for existence of hybrid dacite are phenocrysts that differ from those in rhyolite or basalt and are consistent with equilibrium in an intermediate composition magma. These are augites with composition W037En31Fs31.

In summary, hybrid dacites were formed by mixing of subequal proportions of rhyolite and basaltic andesite. Similar proportions have been found in modeling of hybrid intermediates (e.g., Frost and Mahood, 1987; Feeley and Grunder, 1991) and are controlled mainly by viscosity contrast between mafic and silicic magmas (Sparks and Marshall, 1986; Frost and Mahood, 1987). Any mixing of different magmas is likely to begin with mingling forming "banded" magma which in later stages homogenizes to form one hybrid. In the Rattlesnake Tuff, the existence of hybrid dacites, mingling relics such as resorbed phenocrysts, and discrete and undisturbed banded textures indicate that mixing processes are continuous starting with mingling and followed by hybridization.

Fig. II-12 Dacite mixing models

Average group-E rhyolites (Chapter III) and average of phenocryst-poor inclusion as mixing member. Modeled dacites are #Rt-2A, Rt-141A, and average dacite (see Table II-1 for samples used for average dacite). Ordinate value, X, is required fraction of average group-E rhyolite in mixing model. No mixing solution exists for elements with X of less than zero and greater than 1, because the concentrations of the element are higher or lower in both mixing members. Arrows indicate that value is out of displayed range. Method according to Fourcade & Allegre (1981), with: X = fraction of component a; a = Cm-Cb/Ca-Cb, a = average group-E rhyolite, b = basaltic andesite composite of phenocryst poor inclusion, and m = modeled mixture.



Fig. II-13 Compositional gaps

Gaps between dacites and high-silica rhyolites, symbols as in figure II-2.



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Compositional gap

A compositional gap of 6 weight % SiO₂ exists between least-evolved highsilica rhyolites and dacite composition. No dacite pumice with less than 5 volume % of high-silica rhyolite material falls in the range between 68 and 74 weight % SiO₂. All major and minor elements except the alkalis and aluminum show this gap (Fig. II-2, II-9). Among trace elements, a clear compositional gap is evident only in the compatible elements Sr, Sc, and Co (Fig. II-13). On the other hand, incompatible elements like Rb, Zr(Hf), Nb(Ta), and others have continuous trends from dacite to the high-silica rhyolites (Fig. II-14).

The origin of compositional gaps in silicic magma chambers is highly debated; some favor eruption dynamics (Spera et al., 1986, Blake, 1981, Blake & Ivey, 1986), others favor processes operating in the magma chamber (e.g. Grove & Donnelly-Nolan, 1986, Hildreth, 1987, Bacon & Druitt, 1988, Boden, 1989). In the case of the Rattlesnake Tuff, there is no evidence for the existence of low-silica rhyolite magma. The observed gap in the Rattlesnake Tuff magmas seems real based on: (1) high-silica rhyolite and mafic magmas had to be in physical contact in order to generate hybrid dacites; and (2) the presence of high-silica rhyolite crystals in dacite magma suggests that there was no other magma between dacites and high-silica rhyolites. Rapid crystallization over a narrow temperature interval as suggested by Grove & Donnelly-Nolan (1986) is unlikely because it would require ca. 40% crystallization yielding higher incompatible trace element concentrations in the rhyolites than observed (see 'Rhyolite genesis', this chapter). A gap is observed in elements with a strong elemental gradient between rhyolite and mafic magma, and not where only small element variations between rhyolite and mafic magma exist. Based on the modeled mixing proportions that support rheological arguments for when mixing can occur, we posit that the compositional gap is the result of effective immiscibility between dacite and rhyolite. We further speculate that production of dacite increasingly hindered rhyolitebasalt mixing.

Magma chamber reconstruction

Through a combination of volcanological, textural, compositional, and petrologic lines of evidence, an internally consistent model of the magma chamber that existed just before eruption of the Rattlesnake Tuff can be reconstructed.

The occurrence of only rhyolite pumice in the base of the tuff and in some distal sections, indicates rhyolite erupted first and was at highest level in chamber. The diversity of rhyolites implies there was a compositionally zoned silicic chamber. Banding between dacite and a variety of rhyolites indicates draw up of dacite which is consistent with dacite underlying the rhyolite. Occurrence of dacite proximal to the vent and not in shards indicates that dacite was erupted late consistent with dacite being deeper. Abundant quenched mafic inclusions in the dacite, and not in the rhyolites, indicates that basaltic magmas laid under the dacite. The existence of dacite between rhyolite and enriched basaltic andesite magma is supported by the compositional evidence for mixing and textural and mineral composition evidence for mingling between the two to make dacite. Figure II-15 depicts a magma chamber consistent with the layering of these magma compositions.

Reconstruction of the mafic root zone is more speculative. Enriched mafic magma of grossly basaltic andesite was at the top of the mafic root zone because it mixed to yield dacites. Direct evidence for associated HAOT basalt exists in the form of contemporaneous lavas and mafic inclusions with typical HAOT mineral assemblages within dacite. HAOT can further be inferred to have coexisted, because granular inclusions indicate basalt was crystallizing and because basaltic andesite is derived mainly by fractionation and recharge of HAOT. Therefore, despite the very small volume of mafic constituents in the Rattlesnake Tuff, mafic magma made up the dominant volume of the chamber.

Three lines of evidence suggest that the outlined arrangement was established not only shortly before eruption but existed some time earlier. Differentiation of high-silica rhyolite magma required time as well as high heat input from below. HAOT magma needed time to evolved through fractionation, recharge and possible contamination to the observed enriched and evolved basaltic andesite magma. And thirdly, hybrid dacite magma required time for homogenization after mixing of enriched basaltic andesite with high-silica rhyolite.

Fig. II-14 Scatter plots with petrologic relationships

Schematic interpretation shown with arrows. Symbols for Rattlesnake Tuff samples as in figure II-2, except open diamond for chrysolite & plagioclase inclusion #14Dinc1. Contemporaneous HAOT's are indicated by crosses and andesite #209an shown with open cross. Eu/Eu* calculated with Sm and interpolated Gd* value. F&R is fractionation, recharge, and possibly some assimilation. Fractionation trends within high-silica rhyolites shown with dotted lines.







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The magma chamber was at upper crustal levels based on the consistency of low-pressure models for derivation of basaltic andesite. The best models were those for 0.4 GPa. Because silicic chambers impede the ascent of more dense mafic magma (c.f. Hildreth, 1981, Valentine, 1993), the occurrence of coeval mafic lavas can be interpreted as a lateral limit to silicic caps. Consequently, the location of the contemporaneous HAOT basalts (40 km from the center of the venting area) has to lie outside the edge of silicic zone of the Rattlesnake Tuff chamber.

Conclusions

The Rattlesnake Tuff magma chamber had an extreme compositional diversity of physically contiguous magmas consisting of voluminous high-silica rhyolites (≥ 280 km³), dacites, and basaltic magmas.

Prior to the development of this rhyolite-basalt zoned magma chamber, dehydration partial melting of meta-greywackes or amphibolites is the likeliest process to generate the least evolved high-silica rhyolites. Unusually high Ba/Rb ratios of least evolved high-silica rhyolites require that appropriate source lithologies before melting either contained or developed high Ba/Rb ratios during deep crustal metamorphism.

Evidence from quenched inclusions indicate that the mafic root zone of the Rattlesnake Tuff chamber consisted mainly of trace-element enriched basaltic andesite. Basaltic andesite evolved primarily through fractionation and recharge from typical primitive HAOT which was present at depth. Dacites are a result of mixing between high-silica rhyolites and basaltic andesites (Fig. II-15).

An important aspect of the Rattlesnake magma chamber is that basalt, basaltic andesite, dacite, and rhyolite (itself zoned) coexisted in physical continuity. Once the rhyolite formed, basalts became trapped beneath it and caused a feedback whereby basalt pooled and differentiated and created a compositionally zoned mafic root. Persistence of the mafic root zone in turn thermally maintained the silicic cap thus allowing for interaction at the interface to produce dacites, as well as time for the silicic cap to become zoned. When the thermal input wanes, the cap freezes preventing the continuing formation of dacites and primitive HAOT basalts are no longer trapped and so, no longer produce basaltic andesites. It follows that mafic inclusions, which are exceedingly common in dacites (e.g., Bacon & Druitt, 1988; Feeley & Grunder, 1991) probably represent compositions substantially evolved from a primitive mafic lava whereby more enriched inclusions are thought to be found in systems of greater longevity of the silicic cap, assuming same parental mafic magmas. The Rattlesnake Tuff system displays the critical role of the persistence of a thermal anomaly (injection of basalt) and the importance of a silicic magma in the creation of intermediate composition magmas.

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Chapter III: The Rattlesnake Tuff, part II: Differentiation of the high-silica rhyolites

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Abstract

More than 99% of the 280 km³ of magma erupted as Rattlesnake Ash Flow Tuff, eastern Oregon, is composed of high-silica rhyolite in form of glass shards and pumices. Glassy high-silica rhyolite pumices cluster in 5 chemical groups, three consisting of white pumices and two consisting of dark-gray pumices. Key elements for defining these five high-silica rhyolite compositions are Fe, Ti, La (LREE), Ba, Eu, Rb, Cs, Zr(Hf), Ta(Nb) and Th. On element-element diagrams, the high-silica rhyolite pumice clusters define linear or curvilinear arrays sometimes with gaps between groups. Heavy liquid separation on one representative pumice from each group indicates crystal contents around 1 weight % for the least evolved composition decreasing to 0.09 weight % and to virtually aphyric at the most evolved composition. Progressive changes with higher degree of evolution of the magma occur in mineral modality, mineral chemistry, and partition coefficients. Data on physical parameters indicate that conditions became more oxidizing with differentiation.

Models of crystal fractionation in non-modal proportions of observed phases with empirical partition coefficients agree generally with enrichment trends and yield compositions of removed solids ranging from syenitic to granitic. On the other hand, crystal fractionation of observed phases in modal proportions is incompatible with observed zonation patterns. Fractionation combined with assimilation (AFC) could have occurred during the first 20% (of 60%) fractionation, but is thought to be of minor importance for observed chemical gradients among high-silica rhyolites. Accumulating liquids in a single stratified chamber generated by successive partial melting is inconsistent with extremely low Ba (25 ppm) in most evolved and highest Ba (1900 ppm) in least evolved rhyolite. Derivation of the intermediate high-silica rhyolite compositions through mixing of endmembers is excluded because variable mixing degrees are necessary to explain element abundances of individual pumices.

The likeliest process responsible for the generation of the progressively more evolved high-silica rhyolite compositions is a differentiation process through which each composition is derived from the previous, less evolved liquid. Assuming a slablike chamber geometry, the fractionation process might occur primarily along the roof of the chamber which is the coolest and largest surface. Each successive more evolved and lighter liquid is generated at the roof and stays as the top layer, generating a compositionally and density stratified magma chamber.

Introduction

Single voluminous ash-flow tuffs can be viewed as snapshots in the evolution of large magma chambers. Rapid evacuation and quenching of magmas preserves chemical and physical gradients which existed in the original magma column (e.g. Smith, 1979, Hildreth, 1979, Mahood, 1981, Fridrich & Mahood, 1987, Grunder & Mahood, 1988, Boden, 1989). Rhyolitic ash-flow tuffs commonly have large trace element gradients over small major element variations. Large trace element gradients suggest almost as much fractionation occurs within rhyolites as between basalt and rhyolite when crystal fractionation models are considered. Therefore, the importance of the kind of trace element partitioning process (liquid/crystal, liquid/liquid, gas/liquid) responsible for the main chemical zonation features of felsic magma chambers and the configuration of compositionally zoned magma have been controversial (Hildreth, 1981, Hildreth & Mahood, 1983, Michael, 1983, Miller & Mittlefehldt, 1984).

Other controversial topics concerning zoned ash-flow tuffs include the cause of compositional gaps which are observed frequently. Compositional gaps within silicic tuffs have been ascribed to intrinsic zonation patterns in the magma chamber, such as the vertical stacking of quasi-isolated convection cells (Grove & Donnelly-Nolan, 1986, Fridrich & Mahood, 1987, Hildreth, 1987, Bacon & Druitt, 1988, Boden, 1989), and to eruption dynamics (Blake, 1981, Blake & Ivey, 1986, Spera et al. 1986).

The late Miocene Rattlesnake Tuff from eastern Oregon combines several features making it unique. It is of large volume (280 Km³ dense rock equivalent, DRE), contains abundant glassy pumices which are easily extracted, and has a large chemical zonation. Major element compositions include basalt, basaltic andesite, dacite,

and high-silica rinvolite. Furthermore, high-silica rhyolite is compositionally zoned with large trace element gradients and with compositional clustering. This report focuses on the evolution within high-silica rhyolite magmas of the Rattlesnake Tuff. The generation of the least evolved high-silica rhyolites, dacites, and mafic inclusions are discussed elsewhere (Chapter II).

The Rattlesnake Tuff

Overview and rationale

The 7.05 Ma old Rattlesnake Tuff erupted from the Harney Basin, a center of Late Miocene silicic magmatism in southeastern Oregon (Chapter I). High-silica rhyolites (>75% SiO2) make up more than 99% of ca. 280 km³ DRE of magma erupted. Other eruptive products consist of dacite pumices (≤ 1 volume %) and mafic inclusions (<< 0.01 volume %) (Chapter II). The tuff consists of a single cooling unit, commonly between 10 to 30 m thick and is pumiceous, including pumices as large as 60 cm proximally to the source (Chapter I). Several compositions of rhyolite magma are distinguishable based on clustering of data from glassy pumices and, to a lesser degree, from glass shard populations. Crystal contents of pumices range from less than 2% to virtually aphyric. Therefore observed chemical zonations are almost exclusively features of the liquids. Rhyolites are metaluminous to slightly peralkaline with molar ratios of total alkalis to aluminum of 0.88 to 1.03.

Modeling of silicic magmas requires partition coefficients determined for the system of interest in order for the modeling to be realistic (Mahood & Hildreth, 1983). Partition coefficients in silicic system are variable and high and there is a general lack of experimental data for silicic systems. Therefore, in order to carefully model the relationship between different high-silica rhyolites, partition coefficients for all mineral-bearing high-silica rhyolite clusters were determined on representative samples by analysis of quantitative mineral separates (Table III-1 & III-6). Special care was employed to prepare mineral separates in order to minimize element contributions from trace phase inclusions to bulk mineral analyses (see below).

Analysis and mineral preparation

Sample materials are glassy pumices and glassy shards. Major and trace element data were acquired of whole rock pumices by X-ray fluorescence analysis (XRF) at Stanford University, except #RT220A done at Washington State University, and neutron activation analysis (INAA) at the Oregon State Radiation Center. Uncertainties of analyses are reported in Chapter II.

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Mineral separation on 5 representative pumices were performed. Following crushing with hammer and jaw crusher, pumices were sieved with the following divisions: 2 mm; 1.7 mm; 990 microns; 750 microns; 500 microns; 300 microns; 180 microns; and 106 microns. After first sieving, all material greater than 990 microns was further crushed in small proportions until all pumice material passed through 990 microns mesh size. The total weight of each sample, reported in table III-2, was then measured. Material for glass separates was isolated before heavy-liquid separation. Sequential heavy-liquid (Bromoform and Tetrabromoethane) mineral separations were done on each size fraction with the density of the liquid kept close to the glass densities to ensure that minerals with attached glass also sank. The total mineral yield per sample is expected to be close to the actual mineral content of the pumices because a probable loss of minerals smaller than 106 microns is estimated to be minor and is likely to be compensated by some glass rinds on crystals (Fig. III-1). Total modes of pumices were determined by summation of mineral weights for each size fraction. Some volumetric estimates (recalculated to weight %) were made to determine quartz : feldspar ratios in felsic assemblages and clinopyroxene : magnetite ratios in mafic assemblages. The agreement between weights of total mineral yields before and after separation into individual phases deviated between 1 to 10 % with lower yields in the separated minerals partly attributable to dissolution of glass rinds during wash with dilute hydrofluoric acid after first determination of total mineral yields.

Mineral separates for INAA analysis were prepared by hand-picking from the 300-500 micron fraction for feldspars, 106-300 fraction for magnetites, and mainly 106-180 fraction for pyroxenes. Minerals and glass separates were analyzed as grains as obtained after handpicking. All selected feldspars were clear and inclusion free. Magnetites separates were superficially 100% clean. In case of the pyroxenes, the selected size fraction was small enough so that most pyroxenes were translucent making screening for inclusions feasible. All pyroxene separates were visually clean at > 99%. Mineral separates were multiply washed in mild acids, distilled water, and acetone. Br values of less than 6 ppm for all mineral separates verify the almost

complete removal of heavy liquid residues from mineral surfaces (Table III-4). Glass separates were picked from the 300-500 micron fraction and separates were completely clean of crystals except sample RT165A where very tiny (± 10 microns) Fe-oxides (?) crystals were sparsely but evenly dispersed.

Major and trace element concentrations in mineral and glass separates were determined by instrumental neutron activation analysis (INAA) at the Oregon State University Radiation Center using a 1MW Triga reactor. Weights of analyzed samples ranged from 10 to 70 milligrams. Short activation was performed at a power level of 50 kW for 5 minutes and long activation at 1MW for 12 hours. Counting was done sequentially, three times after short activation and five times after long activation using intrinsic Germanium and low-energy photon (LEP) detectors.

One non-welded glassy bulk tuff sample was used for separation of the different shard populations observed in the matrix of the Rattlesnake Tuff which correspond to the various rhyolite pumice clusters. After loosening shards, magnetic and heavy-liquid techniques were used to separate shard populations. Materials used for analysis were free of crystals and lithic fragments. Trace element concentrations of shard populations were determined by INAA.

Pumice clusters

Definition of clusters

High-silica rhyolite compositions of Rattlesnake Tuff bulk pumices cluster in several groups. The clusters are mainly established by the variation of La(LREE), Eu, Ba, Ta(Nb), Hf(Zr), Rb, Cs, Th, U, Ti and Fe (Fig. III-2). Other elements also vary systematically but the variations are very small. The grouping of the pumices is consistent in most variation plots. The clustering of pumice compositions is largely consistent with clustering of the various shard populations of the matrix; slight deviations between pumice clusters and separated shard populations are most likely due to imperfect separation of shards (Fig. III-2a).

Banded pumices are those with visible signs of different compositions and are listed separately. Samples RT219A, RT62A, both from distal localities, and all fallout pumices are very small pumices (1-3 cm) and were analyzed mainly for correlation purposes.

Fig. III-1 Mineral grain size distribution of pumices

Representative samples of mineral-bearing rhyolites (sample numbers shown). Mineral budgets displayed in < 106 micron window are estimated mineral losses based on extrapolated obtained grain size distribution.



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To evaluate how much trace element concentrations of whole rock pumices are affected by trace elements hosted in minerals, the mineral contributions were calculated using the determined modal abundances and trace element concentrations (Tab. III-2, III-4); for zircon, Zr and Hf from Rattlesnake zircons were used and other element concentrations were taken from the literature. The results are tabulated in table III-5. Trace element contribution from minerals with or without zircon are mostly smaller than the range of analytical uncertainties, except possibly the Ba contribution from anorthoclase in the rhyolite compositions E and D (Table III-5).

Using the grouping established by the scatter diagrams, a mean and one standard deviation was calculated for each group (Table III-1). For most elements, the variance of one standard deviation is in the range of the analytical uncertainty which is sometimes also the difference between adjacent groups causing some overlap. In general, groups A, B, D, and E are well established. Group C has more internal scatter than other groups.

Supporting evidence for the observed chemical clustering comes from the macroscopic color of pumices and shards in the tuff matrix. Group A, B, and C pumices are white, group D pumices are medium gray, and group E pumices are dark gray (see below). Under the microscope, the same gradation is observed among glass shards in the matrix, which range from colorless to dark brown, giving the matrix a "salt & pepper" appearance in hand samples when the tuff is little welded. Microprobe analyses of glass shards from single slides established the positive correlation of Fecontent with darker color of the rhyolitic shards. Vesicularity difference between pumices can be ruled out to explain color gradations because the color variation is evident in pumices at comparable vesicularity and the correspondence with the same color variation in non-welded shards.

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A 1-m thick fallout deposit crops out locally close to the source and is directly overlain by Rattlesnake ignimbrite. This deposit is considered precursory to the main ash-flow eruption based on chemical data between fallout pumices and group A pumices indicating very similar compositions. However, there are some suggestions, mostly through higher Sc, U, Th, Ta, that the fallout pumices represent a magma composition slightly more evolved than group A magma (Table III-1, Fig. III-2a).

To estimate the volumetric proportions of the different rhyolite magma is difficult. A rough estimate can be done by using the distribution of brown versus white shards of the matrix which roughly groups composition A and B and composition D and E. This way, a 1:1 proportion is derived which seems to be a maximum value for the "brown magma" since clear shards dominate distally where the impact by erosion is greatest.

Chemical zonation

Enrichments (or depletions) relative to the average of group E pumices were calculated using the means for each group (fallout pumices excluded) (Fig. III-3). The form of enrichment and depletion is similar to those documented for the Bishop Tuff (Hildreth, 1979), Tala Tuff (Mahood, 1981), and other high-silica rhyolite magma chambers (Hildreth, 1981; Miller & Mittlefehldt, 1984). Elements which are increasingly enriched compared to the least evolved group-E rhyolite average are Si, Cs, Rb, U, Th, Ta, Nb, Pb, Y, HREE, Sb, and possibly Ni; depleted elements are Fe, Ti, Mg, Ca, Ba, Sr, Eu, Zr, Hf, LREE, Zn, and possibly Co. Also, enrichments increase, or depletion decrease, systematically from group E to A with possible minimum or maximum for Sc, Tb, and possibly Al within the intermediate rhyolite groups (Table III-1, Fig. III-2). Conservative elements with no enrichment or depletion are Ga and As. Trace elements with greater uncertainties due to low concentrations are V, Cr, Ni, Co, and Sr. Manganese concentrations are almost constant. More precise Mn data (by INAA) exist only for the glass separates and suggest a minimum in group B. The behavior of Na and K is difficult to evaluate. Na and K concentrations become more variable from group-A to group-E rhyolites (Fig. III-2b), whereas total alkalis (Na₂O+K₂O) are nearly constant with an average of 9.1 weight % and a range from 8.8 to 9.6 weight % without any correlation with Na2O concentrations. Early postemplacement ion exchange is likely to have caused this scattering, increasing K and reducing Na contents (Fisher & Schmincke, 1984, p. 328). This process appears to have taken place without reducing concentrations of total alkalis. The samples richest in sodium are likely to be closest to the original concentration. With this assumption, sodium slightly decreases by ca. 0.6 weight % and potassium increases by the same amount from group-E to group-A. Alkali mobility did not affect other "mobile" elements (Zielinski, 1982) because Na (or K) does not correlate with Cs, Rb, and U (Fig. III-2b). On the other hand, ion exchange affects the calculated agpaitic or

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peralkalinity indices (AI = molar (Na + K)/ Al) (Table III-1) because of higher molar weight of K compared to Na,. The extent of reducing the agpaitic index by ion exchange of K for Na can be hypothetically estimated and is about 0.2 "AI-units" per 10% difference between original Na2O (or K2O) concentration and Na2O (or K2O) concentrations after ion exchange.

Petrography and mineralogy

High-silica rhyolite pumices are crystal poor with crystal contents around 1 volume %. Crystal-enrichment in bulk-tuff compared to pumices due to elutriation of fines (Walker, 1972) is undocumented with the necessary precision needed for such low crystal concentrations. In handsamples, a change in crystal content is recognizable from gray to white pumices with crystal contents dropping from $\pm 1\%$ to virtually aphyric, respectively. and is confirmed by mineral separation data (Table III-2). The increase of crystal contents from 0.6 to 1.3 weight % from group-E to group-C requires verification by more data. No emphasis is put on this increase because it could also be an artifact of pumice sampling. The abrupt drop of crystal content to ca. 0.1 weight % in the group-B pumice (#RT34E) is thought to be significant (Tab. 2). Only ca. 8 crystals (feldspar and quartz) were recovered from group-A pumice (#RT173H).

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Among the crystal-bearing pumices (all but group A), the mode and the mineral chemistry change progressively (Table III-3, Fig. III-4 to III-6). Phases observed in all mineral-bearing pumices are feldspars, Fe-rich clinopyroxene, magnetite, quartz, and the accessory phases zircon, and apatite. Pyrrhotite was found as inclusions in all but in group B pumice. The main modal changes are: (1) the quartz/feldspar ratio increases from a value close to zero to about equal proportions, (2) clinopyroxene and magnetite concentrations are relatively constant, and (3) despite intense search, fayalite is restricted to group-D and -E. In group-D and -E pumice, fayalite and quartz are found in minor quantities (Table III-2). Trace amounts (several flakes) of biotite were found in groups B, D, and E. Some chevkinite inclusions were found in pumice group C (#RT165A) and group E (#RT173C). A few small grains (\approx 100 microns) of high density or magnetic phases (hematite & sulfides ?) were also found and will not be discussed further.

contents are kept mainly intact.

Fig. III-2a:



Fig. III-2b:



Table III-1 Chemical composition of pumices and glass shards Samples are from the Rattlesnake Ash Flow Tuff and from a thin precursor fallout deposit. Major element XRF data are normalized to 100% anhydrous. Prenormalization totals ranged between 98.7 to 100.2%. Analyses are arranged in groups according to clusters observed in scatter diagrams (Fig. III-2). Mean and one sigma standard deviations are calculated. Sample AF1c of fallout pumices is pumice composite of 6 pumices. AI is molar ratio (Na+K)/Al. Numbers in parentheses are anomalously high and are not used, numbers in italic letters indicate values from INAA.

		fallout						
		AF1A	AF1B	AFIC	mean±lo			
INAA	ppm							
FeO*		0.88	0.88	0.88	0.88±0.00			
Na		2.78	2.11	2.30	2.4±0.4			
Cs		3.84	4.45	4.01	4.1±0.3			
U		6.83	5.73	7.39	6.7±0.8			
Th		10.69	11.20	10.80	10.9±0.3			
Hf		8.07	8.15	8.08	8.10±0.04			
Ta		2.57	2.61	2.58	2.59±0.02			
Sb		1.61	1.50	1.48	1.53±0.07			
As		6.1	3.8	4.2	4.7±1.2			
Sc		4.60	4.71	4.52	4.6±0.1			
Co		0.24	0.25	0.17	0.22±0.04			
La		20.9	19.0	17.8	19±2			
Ce		51	53	46	50±4			
Nd		33	32	28	31±3			
Sm		10.58	8.99	8.54	9.4±1.1			
Eu		0.77	0.65	0.56	0.66±0.11			
Tb		2.43	2.19	1.92	2.18±0.26			
Yb		11.26	9.58	8.83	9.9±1			
Lu		1.66	1.50	1.34	1.50±0.16			
Eu/Eu*		0.20	0.19	81.0	0.19±0.01			

			glass	shard	populations			bulk
			separated	from bulk	tuff sample	RT75		tuff
INAA	ppm							RT75
FeO*		0.89	1.16	1.39	1.49	1.79	1.79	1.22
Na		3.49	3.41	3.34	3.32	3.40	3.64	3.51
Cs		3.34	3.11	2.97	2.95	3.01	2.87	3.14
U		4.1	3.7	2.6	3.2	2.4	2.7	3.38
Th		9.16	7.39	6.85	7.02	6.13	5.63	7.39
Hf		8.05	9.36	10.15	10.45	11.30	11.15	9.02
Ta		2.04	1.66	1.46	1.55	1.51	1.32	1.68
Sb		1.75	1.26	1.83	2.59	1.41	1.25	1.32
As		4.1	4.0	4.7	3.59	2.1	4.5	3.7
Sc		4.36	3.76	4.07	4.10	4.64	4.31	4.15
Co		0.21	0.17	0.31	0.31	0.51	0.31	0.23
La		25.6	45.0	51.8	55.6	54.4	55.8	40.1
Ce		70	111	122	135	133	128	94
Nd		35	53	62	67	71	68	48
Sm		10.45	13.25	14.70	14.5	14.23	13.83	12.4
Eu		0.87	1.31	1.66	1.76	2.44	2.42	1.45
Tb		2.39	2.51	2.32	2.39	2.17	2.07	2.26
Yb		9.99	9.57	9.02	8.84	8.36	8.17	9.26
Lu		1.51	1.42	1.36	1.35	1.26	1.25	1.38
Eu/Eu	*	0.22	0.28	0.35	0.37	0.53	0.54	0.34

Table III-1, Continued:

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		group-A						
	RT14F	RT55B	RT165E	RT173D	RT173E	RT173H	mean±lσ	
SiO2	76.00	77.44	77.60	77.77	77.82	77.67	77.4 ±0.7	
TiO2	0.12	0.11	0.11	0.12	0.11	0.11	0.11±0.01	
A12O3	11.72	12.10	12.07	11.97	11.91	11.96	11.9±0.1	
Fe2O3*	0.86	0.86	0.86	0.89	0.87	0.87	0.87±0.01	
MnO	0.08	0.07	0.08	0.09	0.08	0.08	0.08±0.01	
MgO	0.10	0.13	n.d.	0.03	0.04	0.05	0.06±0.05	
CaO	(2.32)	0.30	0.30	0.28	0.26	0.26	0.28±0.02	
Na2O	3.33	3.35	3.89	3.68	3.81	3.46	3.59±0.24	
K2O	5.47	5.62	5.06	5.17	5.08	5.53	5.32±0.25	
P2O5	n.d.	n.d.	0.02	0.01	0.02	0.02	0.01±0.1	
AI	0.97	0.96	0.98	0.97	0.99	0.98	0.98±0.01	
XRF ppm								
Rb	116	121	122.7	123.1	122	122.4	121.2±2.7	
Ba	19	40	24.6	17.7	9.6	39	25±12	
Sr	(36.8)	3.3	3.4	2	1.1	2.1	2±1	
Zr	169	180	175.3	173.6	174.9	174.7	175±4	
Nb	37.4	39	40.3	40.6	39.3	39.4	39±1	
Y	96.4	98	100.5	100.7	99.8	100.3	99±2	
Pb	17.9	18.3	18.9	21.1	17.2	20	19±1	
Zn	53	79	88.2	90.5	86.9	87.5	81±14	
Ga	17.2	18.4	17.8	17.8	17.9	17.6	17.8±0.4	
v	6.4	2.4	2.5	n.d.	n.d.	n.d.	4±2	
Cu	1.9	0.8	1.2	2	1.4	1	1.4±0.5	
Ni	10.9	11.1	14	12.6	12.6	14.3	13±1	
Cr	1.5	2.5	1.6	n.d.	n.d.	n.d.	1.9±0.6	
INAA ppu	u 0.75	0 74	0.8	0.76	0 78	0 78	0 77+0 02	
reu*	3.4	3 11	3.86	3.6	3 62	3 3 5	3 5+0 3	
Na Ca	5.4	4 22	J.80 A 54	4 28	4 5	4 4 8	4 4+0 2	
	4.1	4.23	4.34	4.20	4.5	4.40	4 7+0 2	
0 Th	9.01	4.05	0.87	9.79	9.62	9 47	9 5+0 4	
	6.00	5.75	7 24	7 1 2	7 15	7 13	7 1+0 2	
	2.05	2.15	2 28	2 14	2 15	2 16	2 16+0 07	
1a Ch	1.30	1 43	1 4 2	1.5	1.52	1.55	1 46+0 08	
30	1.52	2 5 2	5 31	3 87	4 38	4 38	4 4+0 6	
AS 5 a	2 76	3.06	4.05	3.0	3 80	3 93	3 9+0 1	
Co	0.42	0.08	0.05	0.06	0.05	0.08	0.1±0.2	
1 a	19.9	19.1	21.7	20	19	19.9	20±1	
Ce	46.4	48.5	54	5 5	51	49	51±3	
Nd	29.1	23.4	30	31	3 2	28	29±3	
Sm	9.09	8.83	9.58	9.4	9.06	9.57	9.3±0.3	
En	0.65	0.67	0.65	0.64	0.65	0.65	0.65±0.01	
Tb	2.18	2.15	2.3	2.2	2.17	2.22	2.20±0.05	
Yb	9.5	9.62	10.55	10.4	10.2	10.54	10.1±0.5	
Lu	1.45	1.49	1.62	1.62	1.54	1.61	1.56±0.07	
Eu/Eu*	0.19	0.20	0.18	0.18	0.19	0.18	0.19±0.01	

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	RT34D	RT34E	RT120A	RT220A	RT219A	mean ±1σ
SiO2	76.53	77.18	77.02	77.06		77.0±0.3
TiO2	0.12	0.12	0.11	0.12		0.12 ±0.01
A12O3	11.82	11.77	11.80	11.82		11.8 ±0.02
Fe2O3*	1.32	1.37	1.41	1.39		1.37 ±0.04
MnO	0.09	0.09	0.09	0.10		0.09 ±0.01
MgO	0.10	0.09	0.02	0.08		0.07 ± 0.04
CaO	0.35	0.33	0.39	0.26		0.33 ±0.06
Na2O	3.19	3.06	3.36	3.72		3.3 ±0.3
K20	6.43	5.95	5.74	5.40		5.9 ±0.4
P2O5	0.05	0.03	0.05	0.05		0.05 ±0.01
AI	1.03	0.97	1.00	1.01		1.0 ±0.3
XRF ppm						
Rb	91	91	93.6	94		92 ±2
Ba	122	132	130.8	115		125 ±8
Sr	5	3.7	12.5	12		8 ±5
Zr	304	308	301.6	258		290 ±23
Nb	31.5	32	33	32		32.1 ±0.6
Y	93	96	98	89		94 ±4
Pb	18	19.3	16.9	19		18 ±2
Zn	99	105	101	101		101 ± 3
Ga	17.1	17.9	18	21		19 ±2
V	5.6	5.8	1.4	n.d.		4 ±2
Cu	2.1	1.9	4.6	7		4 ±2
Ni	9.8	9.5	15	14		12 ±3
Cr	2.2	2.4	0.6	2		1.8 ±0.8
INAA ppr	n					
FeO*	1.10	1.20	1.29	1.24	1.27	1.22 ±0.08
Na	3.53	3.42	3.30	3.44	1.94	3.1 ±0.7
Cs	3.07	3.19	3.30	3.18	2.88	3.1 ±0.2
U	3.26	3.52	3.62	3.63	3.17	3.4 ±0.2
Th	7.15	7.71	7.69	7.84	7.57	7.6 ±0.3
Hf	9.09	9.61	9.41	9.68	9.79	9.5 ±0.3
Та	1.68	1.75	1.81	1.71	1.62	1.71 ±0.07
Sb	1.26	1.33	1.41	1.39	1.7	1.4 ± 0.2
As	4.72	n.d.	5.74	4.5	5.3	5.1±0.6
Sc	3.47	3.69	3.84	3.77	3.97	3.8 ± 0.2
Co	0.18	0.32	0.28	0.14	0.58	0.3 ±0.2
La	37.4	38.9	39.4	38.7	37.5	38 ±1
Ce	84.5	89.7	97	99	87.8	92 ±6
Nd	46.8	44.9	53	50	48.1	49 ±3
Sm	12.63	12.89	13.57	13.45	12.11	12.9 ±0.6
Eu	1.07	1.22	1.16	1.21	1.27	1.19 ±0.08
Tb	2.53	2.7	2.68	2.64	2.14	2.54 ±0.23
Yb	9.48	9.66	9.79	9.61	8.48	9.4 ±0.5
Lu	1.35	1.38	1.42	1.37	1.23	1.35 ±0.07
Eu/Eu*	0.24	0.26	0.24	0.26	0.31	0.26 ±0.03
Table III-1, Continued:

			group-	С			banded	
0.00	RT165A	RT120B	RT80A	RT62A	mean ±1σ	RT17A	RT55C	RT127A
SiO2	77.11	76.82	76.63		76.9 ±0.2	75.98	76.24	75.23
TiO2	0.12	0.13	0.14		0.13 ±0.01	0.14	0.15	0.16
A12O3	11.77	11.67	12.06		11.8 ±0.2	12.23	12.07	12.11
Fe2O3*	1.61	1.64	1.91		1.72 ±0.2	1.67	1.87	2.09
MnO	0.09	0.08	0.08		0.08 ±0.01	0.09	0.09	0.09
MgO	n.d.	n.d.	0.04			0.07	0.13	0.10
CaO	0.35	0.43	0.32		0.37 ±0.06	0.87	0.43	0.69
Na2O	3.70	3.74	3.01		3.5 ±0.4	2.72	2.22	2.84
K20	5.23	5.43	5.81		5.5 ±0.3	6.20	6.77	6.62
P2O5	0.01	0.06	n.d.			0.03	0.01	0.06
AI	1.00	1.03	0.93		0.99 ±0.05	0.91	0.91	0.98
XRF ppm								
Rb	80	77.6	77		78 ±2	84	77.3	70
Ba	379	491.4	664		511 ±144	843	997.5	1825
Sr	6.5	12.5	6.8		9 ±3	19	13.6	25
Zr	371	372.3	404		382 ±19	355	395.3	444
Nb	30.5	29	28.3		29 ±1	29.9	27.7	26.8
Y	89	8 5	8 2		85 ±3	86	79.1	73.7
Pb	16.4	13.6	17.4		16 ±2	8.9	15.4	15.9
Zn	106	102.8	113		107 ±5	60	114.6	116
Ga	17.7	18.1	18.2		18.0 ±0.3	17.9	17.8	17.5
V	1.5	0.1	2.9		1.5 ±1	2.1	1.9	4
Cu	1.9	4.2	2.9		3 ±1	3.3	8.9	4.3
Ni	9.9	11.6	10.1		11 ± 1	9.3	10.6	10
Cr INAA nom	0.9		1.3		1.1 ±0.3	3.1	4	n.d.
FeO*	1 27	1 44	1 69	1.44	1 40 10 14	1.40		
Na	7.69	2 62	1.00	1.44	1.48 ±0.14	1.48	1.71	1.82
Ca	2.00	3.02	2.70	3.08	3.3 ±0.4	2.96	2.28	2.79
US II	2.0	2.04	2.71	3.03	2.8 ±0.2	2.96	2.63	2.43
Th	5.12	2.07	2.80	3.18	3.0 ±0.2	3.5	2.6	2.7
1 II Uf	0.07	0.80	0.93	7.94	7.1 ±0.5	7.11	6.72	5.64
	9.80	10.04	10.73	10.5	10.3 ±0.4	10.05	10.41	10.63
C L	1.30	1.52	1.51	1.81	1.60 ± 0.14	1.62	1.53	1.41
30	1.10	1.17	1.28	1.33	1.2 ±0.1	1.28	1.22	1.19
AS So	2.55	4.5	n.d.	4.8	4.8 ±0.4	4	3.9	3.5
Co	0.08	0.17	0.39	4.34	3.9 ± 0.4 0.3 ±0.2	4.07	4.85	4.18
I.a.	10.0	51 7	50.4	45.1	10.12			
Ca	49.9	124	50.4	45.1	49 ±3	44.5	49.9	50.4
Nd	5.0	124	111.5	107.5	114 ±/	101.6	93	125
Sm	14 55	14.44	49.4	59.4	57 ±5	48	56	4 5
En	14.55	14.44	12.20	14.0	14.0 ±1.1	13	14.02	13.41
Th	2.47	1.54	1.81	1.34	1.54 ±0.20	1.71	1.94	2.26
Vh	2.42	2.32	2.39	2.04	2.44 ±0.14	2.33	2.22	2.06
In	0.38	0.38	8.22	9.98	8.8 ±0.8	8.89	8.24	7.91
En/En*	0.20	1.33	1.23	1.45	1.32 ±0.09	1.32	1.25	1.19
Cu/Cu-	0.30	0.32	0.42	0.27	0.33 ±0.07	0.39	0.42	0.52

Table III-1, Continued:

			group-D						
6:00		RT50D	RT173B	RT1731	RT173L	RT4A	mean +1 -		
5102		76.12	76.59			75.31			
1102	-	0.14	0.14			0.14	0.14 +0.00		
A120.	, ,	11.88	I1.97			12.17	12 0 +0 2		
Fe2U	5-	1.81	1.93			1.95	1 89 +0 07		
MnO		0.08	0.09			0.09			
MgO		0.06	0.05			(0.50)	0.06 +0.01		
		0.62	0.45			0.65	0.00 ± 0.01		
Na2O		2.97	3.84			2.35	31+0		
R20		6.29	4.92			6.83	6 02 +1		
P205		0.03	0.02			0.01	0.02 ±0.01		
AI		0.99	0.97			0.02	0.04.10.00		
XRF	ppm					0.93	0.96 ± 0.03		
Rb		67	66.7	63	68	80.0	(0)		
Ba		1201	1230.2	1120	1140	1114	69 ±7		
Sr		18.8	12.4		1140	24.2	1160 ±51		
Zr		426	433			422	19 ±6		
Nb		26.0	27.2			432	430 ±4		
Y		75.5	79.0			23.0	26.3.±0.8		
Pb		16.1	14.8			/0.0	75±4		
Zn		106	115.1			10.9	16±1		
Ga		18	17.6			107	109 ±5		
v		4.2	1.8			17.9	17.8 ±0.2		
Cu		3.3	4			3.7	3±1		
Ni		9.1	9.7			0.3	5±2		
Сг		2.5	n.d.			0.4	9.0±0.7		
INAA	ppm					4			
FeO*		1.57	1.69	1.69	1 7 2	1 71	1 60 10 06		
Na		2.85	3.93	3.28	3 47	2.25	1.08 ±0.06		
Cs		2.27	2.38	2.17	2.41	2.35	3.2 ±0.6		
U		2.41	3.16	2.49	2 44	J.4 1 2 2	2.5 ±0.5		
Th		6.4	6.47	5.87	6.02	2.35	2.6 ±0.3		
Hf		10.39	10.55	10.3	10.6	0.15	6.2 ±0.3		
Та		1.34	1.34	1.29	1 3 8	10.72	10.5 ±0.2		
Sb		1.16	1.19	1.14	1.50	1.4	1.35 ±0.04		
As			3.9	3.7	3 1 3	3.6	1.13 ±0.05		
Sc		3.57	3.76	3.65	3 57	3.72	3.0 ±0.3		
Co		0.4	0.11	0.12	0.24	0.26	0.2 ± 0.1		
La		53	54 3	51	52.0	40.0			
Ce		117.2	130	117	110	49.3	52 ±2		
Nd		52.9	68	57.6	59.0	113	119 ±6		
Sm		12.16	14.01	13 1	12 0	50.8	58 ±7		
Eu		1.97	2.03	2 01	13.0	12.42	13.1 ±0.8		
Тb		2.17	2.23	2.01	2.04	1.94	2.00 ±0.04		
Yb		7.9	7.95	7 32	4.10 7 59	2.00	2.15 ± 0.06		
Lu		1.2	1.22	1.08	1.30	1.15	7.6 ±0.4		
Eu/Eu*		0.48	0.44	0 4 6	1.13	1.07	1.14 ±0.07		
		-		0.40	0.43	0.47	0.46 ±0.01		

Table III-1, Continued:

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	group-E							
	RT34A	RT34B	RT34C	RT55A	RT140A	RT173A	RT173C	mean ±1σ
SiO2	75.32	75.42	75.08	75.73	74.52	75.70	75.47	75.3 ±0.4
TiO2	0.17	0.18	0.18	0.16	0.18	0.16	0.16	0.17 ±0.01
A12O3	12.42	12.19	12.32	12.19	13.04	12.47	12.26	12.4 ±0.3
Fe2O3*	2.23	2.07	2.34	2.07	2.37	2.23	2.18	2.21 ±0.12
MnO	0.09	0.07	0.09	0.07	0.11	0.10	0.10	0.09 ±0.02
MgO	0.12	0.15	0.07	0.08	0.10	n.d.	0.01	0.08 ±0.06
CaO	0.49	0.55	0.63	0.63	0.37	0.54	0.50	0.53 ±0.09
Na2O	3.57	3.56	3.88	2.64	2.54	4.18	4.61	3.6 ±0.8
K2O	5.57	5.77	5.38	6.34	6.75	4.59	4.69	5.6 ±0.8
P2O5	0.02	0.04	0.03	0.08	n.d.	0.02	0.01	0.03 ±0.02
Al	0.96	0.99	0.99	0.92	0.88	0.95	1.03	0.96 ±0.05
XRF ppm								
Rb	63	64	63	62	71	62.2	63.5	64 ±3
Ba	1947	1898	1914	1999	1839	2030.6	1834.6	1920 ±75
Sr	23.9	22.8	26.3	29	18	22.5	23.4	24 ±3
Zr	457	469	460	464	488	474.3	457	467 ±1 1
Nb	25.6	25.2	25.4	25.8	26.7	26.8	26.4	26.0 ± 0.6
Y	76	75	75	74	78	77	76.2	76 ±1
Pb	15.4	14.7	14.9	15.4	9.7	13.1	10.3	13 ±2
Zn	117	106	117	108	110	119.8	113.1	113 ±5
Ga	19	18.3	18.3	18.3	19.9	19	18.6	18.8 ± 0.6
v	7.7	20.9	16.8	4.4	n.d.	3.7	1.9	9 ±8
Cu	3.1	4.1	3	3.5	8.3	4.3	4.7	4 ±2
Ni	8.4	10.6	9.4	8.2	10.7	10	8.8	9 ±1
Cr	4.2	3.4	2.9	1	n.d.	0.3	0.3	2±2
INAA ppm								
FeO*	2	1.87	2.09	1.83	2.00	2.00	1.91	1.96 ±0.09
Na	3.86	3.9	3.71	2.48	2.42	4.04	4.5	3.6 ±0.8
Cs	2.28	2.32	2.33	2.25	2.44	2.33	2.35	2.33 ± 0.06
U	2.3	2.44	2.35	3.1	2.41	2.68	2.18	2.5 ± 0.3
Th	5.63	5.69	6.12	5.47	5.33	5.7	5.67	5.7 ±0.2
Hſ	11.17	11.31	11.19	11.11	10.76	11.2	11.07	11.1 ±0.2
Ta	1.33	1.36	1.35	1.34	1.37	1.36	1.35	1.35 ± 0.01
Sb	1.18	1.28	1.34	1.08	1.16	1.14	1.17	1.19 ±0.09
As	4.9	5.4	4.5	3.4	4.3	4.2	4	4.4 ±0.6
Sc	4.29	4.33	4.83	4.12	5.02	4.51	4.32	4.5 ± 0.3
Co	0.26	0.5	0.59	0.5	0.22	0.14	0.13	0.3 ± 0.2
La	51.1	51	49	49.8	53.5	51.9	51.7	51 ±1
Ce	111.8	109.9	105.6	107.8	126	120	112	113 ±/
Nd	52	50.4	50.5	49.9	0.5	01	38	12 0 10 0
Sm	12.85	12.76	12.26	11.75	14.16	13.30	13.39	13.0 IU.8
Eu	2.56	2.95	2.71	2.5	2.67	2.0	2.48	2.04 ±0.10
ТЪ	2.17	2.22	2.16	2.11	2.12	2.10	2.13	2.15 ±0.04
Yb	7.68	7.65	7.51	7.56	7.99	7.97	7.91	7.8 ±0.2
Lu	1.16	1.14	1.08	1.16	1.25	1.24	1.23	1.18 ±0.06
Eu/Eu*	0.60	0.68	0.65	0.62	0.59	0.58	0.50	0.61 ±0.04

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Table III-2 Modal and absolute mineral abundances

Representative samples of all pumices clusters, derived by heavy liquid mineral separation. Pumice from cluster A (#RT173H) had a total weight of 580 g, but only a few mineral grains were retrieved making the pumice "aphyric".

	Heavy liquid	mineral	modes	
sample	RT34E	RT165A	RT173B	RT173C
pumice grp.	В	С	D	E
minerals:		weight	%	
alkali-fsp	0.025	0.736	0.874	0.522
quartz	0.032	0.536	0.022	0.002
CDX	0.007	0.028	0.055	0.061
magnetite	0.027	0.026	0.041	0.043
favalite			0.001	0.007
biotite	trace		trace	trace
total min.%	0.091	1.326	0.992	0.635
glass:	99.910	98.674	99.008	99.364
total sample				
weight [g]	477	1022	752	1154
accessory	minerals:	(X = preset	ent)	
zircon	X	X	Х	X
apatite	Х	X	X	X
pyrrhotite		X	X	Х
chevkinite		Х		Х

Fig. III-3 Enrichment factors

Enrichment of the average composition of individual pumice clusters relative to the average of least evolved composition rhyolite group E (Table III-1).



Accessory phases occur only as inclusions in major phases, free accessory phases were not observed in even the smallest size fractions except for sparse zircons recovered from the 106-180 micron fraction of group B pumice (#RT34E). Petrographically, accessory phases occur as inclusions in or adhering to clinopyroxene, fayalite, and magnetite grains; inclusions in feldspar are rare. Crystallization of accessory phases seems to be controlled by local saturation around larger, mainly mafic phases (c.f. Bacon, 1989). Zircon is the most abundant phase with less apatite and much less chevkinite; the estimated proportions of zircon to apatite to chevkinite are 10:3:1. The estimate for chevkinite is a maximum value because only 3 grains were found in two pumices compared to more than 30 grains of zircon inclusions per mineral-bearing pumice. An estimate of the absolute concentrations of zircon can be made by determining the volume % zircon in mafic phases and combined with absolute abundances of mafic phases (Table III-2). The volume % zircon in mafic phases was determined by petrographic inspection of mafic phases under reflected light followed by areal analysis of one large clinopyroxene grain (1mm) which appeared to contain more zircons than most clinopyroxenes, fayalites, and magnetites. The volume % of zircon of this grain is 0.93; assuming the same overall zircon abundance in mafic phases, the recalculated weight % of zircon in mafic phases is 1% (4.0 g-cm⁻³ used for "bulk mafic minerals"). From this follows that the total abundance of zircon varies from 0.0005-6 weight % in pumice of group B and C to 0.001 weight % in pumice of group D and E.

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Felsic minerals

Feldspar change from anorthoclase in group-E and group-D pumice to Nasanidine in group-C and -B (Fig. III-4). A few micro feldspar phenocrysts found in sample group-A pumice (#RT173H) are oligoclases (Ab₆₉₋₇₀); however whether they are true equilibrium phenocrysts is ambiguous due to the low abundance. Individual feldspar phenocrysts in all high-silica rhyolite compositions indicate a very limited internal range, commonly with less than 1 Ab unit, except some feldspars from one banded pumice (#RT17A) which are anorthoclases (Ab₆₆) with a more sodic rim (Ab₇₅).

The range of feldspar compositions becomes progressively tighter with 6 Ab units in group-C rhyolite and a 2 Ab unit range in group-B rhyolite (Fig. III-4). The general decrease in Na and increase in K is confirmed by the bulk feldspar analyses

(Table III-4). Ca and Ba concentrations progressively decrease, with BaO concentrations commonly twice those of CaO.

Quartz is bipyramidal and crystal faces indicate only incipient resorption. Quartz is the most common mineral in group-B and C pumice (Tab. III-2). In group D and E, quartz is scarce.

Pyroxene

Pyroxene in the high-silica rhyolites range in composition from Fe-hedenbergite to Fe-augite (Table III-3, Fig. III-5). Individual grains are unzoned with ranges typically less than 1% ferrosilite component. Clinopyroxenes from the samples representing each rhyolite group (group-A excluded) indicate overlap between each adjacent group, but generally indicate that pyroxenes become progressively less-Fe rich, at constant Ca, with higher degree of evolution of the liquid (Fig. III-5). Along with the decrease in FeO*, MnO concentrations increase from around 2.3 to 3.0 weight % with excellent agreement between microprobe and INAA data (Table III-4, Fig. III-5). Bulk mineral analyses verify the progressive trends of decreasing FeO* and increasing MgO and MnO concentrations (Table III-4). Mg and Ca concentrations of bulk pumices are generally within analytical errors of INAA and microprobe analysis (Chapter II).

Fayalite

Fayalites occur only in the two least evolved rhyolite compositions. The compositional range of individual phenocrysts and within pumice clasts is no more than is 0.1 Fo units (Table III-3). Fayalites in group D have slightly, but likely significantly, higher concentrations of MgO and MnO. Therefore magnesium and manganese concentrations increase with higher evolution degree from group-E to -D composition. Manganese concentration of 4.8 to 5.5 wt% are higher than manganese concentrations of fayalites from other Harney Basin rhyolites (MacLean, 1994), from metaluminous to slightly peralkaline high-silica rhyolites (Mahood, 1981), from silicic glasses from Iceland (Carmichael, 1960), and are higher than any reported manganese concentration of fayalites found in Deer et al. (1982).

Table III-3 Representative microprobe analyses of minerals HSRgrp : rhyolite cluster to which mineral analyses belong. Mineral identification includes whether analysis was done on the center of the grain, ctr, at the rim, rim, or somewhere in between, hlf. *: range is for three andesine grains. Fe₂O₃ by charge balance. Or=KAlSi₃O₈, Ab=NaAlSi₃O₈, An=CaAl₂Si₂O₈, Wo=Ca₂Si₂O₆, En=Mg₂Si₂O₆, Fs=(Fe, Mn)₂Si₂O₆, Fo=Mg₂SiO₄, Fa=Fe₂SiO₄, Te=Mn₂SiO₄. For zircons, n indicates number of averaged analyses and errors are one sigma standard deviations; analyses were normalized to 100% before calculating averages and standard deviations.

	feldspar							
HSRgrp	A	B	С	D	E			
sample	RT173H	RT34E	RT165A	RT173B	RT173C			
ID#	B.rim	B.ctr	A.hlf	D.rim	C.rim			
SiO2	61.56	65.52	65.71	65.07	65.36			
A12O3	23.67	19.27	19.30	20.12	20.28			
FeO	0.27	0.28	0.25	0.20	0.22			
MgO	0.01	n.d.	n.d.	n.d.	n.d.			
CaO	4.91	0.18	0.28	0.63	0.68			
Na2O	7.96	5.89	6.24	7.52	7.63			
K20	1.02	7.91	7.08	4.66	4.14			
BaO	0.22	0.27	0.70	1.40	1.26			
total	99.63	99.32	99.56	99.59	99.56			
Or	5.9	46.5	42.2	28.1	25.4			
Ab	70.1	52.6	56.4	68.8	71.1			
An	24.0	0.9	1.4	3.2	3.5			
Ab range					(1 1 75			
	*69-70	51.5-52.8	54-60	67.6-73.7	04.4-75			

	clinopyroxene							
HSRgrp	В	С	D	E				
sample	RT34E	RT165A	RT173B	RT173C				
ID#	I.hlf	A.ctr	1.hlf	G.hlf				
SiO2	49.57	49.1	47.84	48.16				
A12O3	0.37	0.34	0.37	0.33				
FeO	20.44	24.13	27.75	26.98				
MnO	3.03	2.74	2.4	2.34				
TiO2	0.17	0.18	0.26	0.23				
CaO	18.51	18.28	17.59	17.84				
Na2O	0.56	0.53	0.54	0.5				
total	99.14	99.92	99.34	99.45				
Wo	41.74	41.5	41.02	41.2				
En	19.65	13.98	7.67	9.19				
Fs	38.61	44.52	51.31	49.61				
Fs range				10.50.0				
	37.8-40.3	42-48.5	44.5-52	48-52.8				

Table III-3, Continued:

	magnetite						
HSRgrp	В	С	D	E			
sample	RT34E	RT165A	RT173B	RT173C			
ID#	A.ctr	H.ctr	G.hlf	C.ctr			
SiO2	0.07	0.10	0.09	0.10			
TiO2	12.04	15.51	18.87	20.34			
A12O3	0.47	0.43	0.48	0.56			
V2O5	0.08	0.11	0.08	0.12			
Fe2O3	44.82	38.65	31.43	28.45			
FeO	39.71	43.39	46.51	47.92			
MnO	2.07	1.93	1.76	1.72			
MgO	0.20	0.15	0.07	0.07			
ZnO	0.47	0.42	0.38	0.35			
total	99.93	100.69	99.67	99.63			
TiO2 range	11.8-12.6	15.2-15.7	18.6-19.2	19-21.4			

	fay	alite			biotite	
HSRgrp	D	E		B	D	F
sample	RT173B	RT173C		RT34E	RT173B	RT173C
ID#	D.hlf	H.hlf		C.pt2	C.pt2	C nt2
SiO2	29.85	29.35	SiO2	35.35	37.56	35 57
TiO2	0.03	0.04	TiO2	2.40	2.47	2 53
A12O3	0.01	0.02	A12O3	13.40	13.95	13 70
FeO	63.18	63.74	FeO*	25.01	17.02	25.27
MnO	5.59	4.85	MnO	0.25	0.59	0.24
MgO	1.50	1.28	MgO	7.91	13.25	8 18
CaO	0.29	0.35	CaO	n.d.	0.01	n.d
NiO	n.d.	0.03	Na2O	0.37	0.07	0.13
ZnO	0.20	0.16	K20	9.02	9.59	9.55
			H2O	3.04	3.39	2.85
			F	1.45	1.13	1.96
5.000			CI	0.05	0.02	0.04
total	100.65	99.82	total	98.25	99.11	100 11
-			O = F, Cl	0.63	0.48	0.84
FO	3.7	3.2	total	97.62	98.63	99.27
Fa	88.4	90				
le	7.9	6.9	ZnO¥	0.16	0.16	0.05
			BaO¥	0.02	0.63	0.04
Fo range	3.7-3.8	3.1-3.2	MgO range		12.4-13.2	7.9-8.2

Table III-3, Continued:

Ce2O3 Sm2O3 Yb2O3 Y2O3 P2O5

total O=F, Cl

total

F

CI

F
1730
13
1+0 27
+0.03
+0.11
+0.03
± 0.03
+0.02
+0.05
+0.03
±0.07
)
73B
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30
0.0
0.0
50
39

3.58

0.79

0.11

1.87 33.67

2.61

0.22

96.38

1.15

95.23

host= mineral by which analyzed apatite was enclosed

49.38 2.22

0.43

< 0.0

0.87 36.70

2.87 0.22 98.77

1.26

97.52

55.39

< 0.0

< 0.0

< 0.0 < 0.0 41.50

3.57

< 0.0 101.55 1.50 100.05

2.28

0.70 0.19

1.96 36.56 3.81 0.21 98.13

1.65

96.47

Fig. III-4 Feldspar compositions

Each symbol represents one analysis. Capital letters indicate high-silica rhyolite group. Tie-line to hypothetical sanidine composition corresponding to plagioclases Ab₇₀ in sample #RT173H were drawn with information from (1988) at 1kbar and 750°C. Insert shows all feldspar (except plagioclases from sample RT173H) in relation to single feldspar field at 1kb, and 825°C from Fuhrman & Lindsley (1988). Capital letters indicate high-silica rhyolite compositions.



Fig. III-5 Clinopyroxene and fayalite compositions

Top: Pyroxene compositions projected onto the pyroxene quadrilateral and olivine compositions projected onto the Mg - Fe*+Mn tieline; each symbol represents one analysis; all compositions in molar quantities. Capital letters indicate high-silica rhyolite groups.

Bottom: Variation diagrams of clinopyroxenes and fayalites.



Mg

Fe*+Mn



Fig. III-6 Magnetite compositions

Compositions in weight % projected onto a FeO-Fe₂O₃-TiO₂ ternary plot, showing the x-y tie line. Endmembers of magnetite series are: Usp=ulvöspinel, Mag=magnetite. Capital letters indicate high-silica rhyolite groups.



Magnetite

Magnetite is the only FeTi-oxide found in Rattlesnake Tuff high-silica rhyolites. Magnetite compositions become progressively poorer in Ti from group-E pumice to group-B pumice (Table III-3). There is some overlap in the range of Ti concentrations of single crystals between the two least evolved compositions. A compositional gap of Ti in magnetites develops towards the more evolved rhyolite compositions (Table III-3, Fig. III-6). Bulk mineral analyses confirm nicely the progressive decrease in Ti and small increase in manganese with degree of evolution.

Biotite

Only trace amounts (several flakes) of biotite were found in group B, D, and E (Table III-3). Arguments favoring a cognate magmatic origin are the relative large size of flakes (about 0.5 to 2 mm) and the unaltered appearance. On the other hand, euhedral apatite inclusions in biotite are distinctly different from apatite inclusions in other mafic phenocrysts of the same pumice clast; apatite inclusions in magnetite contain significantly more SiO2, FeO, REE and less CaO, P2O5, F (Table III-2). Magnesium and manganese contents of the biotites increase while FeO* decreases from group-E to group-D pumice. Biotites from sample #RT34E indicate reverse trends. Fluorine contents of all biotites are high; most are between 1 to 2 wt.%.

Estimation of intensive parameters

Temperature

Single feldspars can be used to estimate minimum temperatures for compositions B through E. All feldspars from composition B through E plot on or above the line dividing the single from the two feldspar field at 750°C, 1kb and below the dividing line at 825°C, 1kb of the models by Fuhrman & Lindsley (1988) (Fig. III-4). Therefore, for compositions B to E, minimum temperatures are ca. 800 to 830°C at 1kb pressure. The relatively high minimum temperatures are consistent with other fayalite-bearing rhyolites (Carmichael, 1967, Honjo et al., 1992). An upper temperature bracket is given by plagioclases of some of the mafic inclusions incorporated into dacite pumices (Chapter II). Quench textures of phenocrysts suggest a delta-T of ca. 200°C existed between basaltic and dacitic magmas, based on comparison to experimental data by Lofgren (1980). Olivineplagioclase assemblages of the quenched inclusions indicate modeled crystallization temperatures between 1170 to 1270 °C at the given composition using the program COMAGMAT (Ariskin, 1992, Chapter II). Therefore, a upper limit of 970 to 1070°C of the hottest rhyolite in contact with dacite seems reasonable when a small temperature gradient between dacite and rhyolite magmas exists.

Pressure and water fugacity

The position of the pumice data in the Qz-Ab-Or ternary can be used to estimate confining pressures (Fig. III-7). Only the samples with the highest Na/K ratios (least hydrated) for each group are considered. Data from compositions A to D indicate ca. 1 kb and group E rhyolites slightly higher values. These estimates are based on water-saturated conditions. Under water-undersaturated conditions, minima shift horizontally towards the Or-SiO₂ tieline (Johannes & Holz, 1990). Therefore, pressure estimates between water-saturated to -undersaturated conditions should not be much different. The general conclusion is that all high-silica rhyolites indicate upper crustal levels and that less evolved high-silica rhyolites indicate slightly deeper parts.

Progressive increase in potassium in the feldspars crystallizing under hypersolvus conditions (group-E to -B) indicates that coexisting liquids became progressively depleted in Na (Fig. III-4) in accordance with the chemical record (Table III-1); the assemblage feldspar and euhedral quartz indicates cotectic crystallization of both phases. Normative pumice data fall near the 1 kb quartz-feldspar cotectic, but to the right of the minimum in the Qz-Ab-Or ternary (Fig. III-7). Liquids to the right of the minimum should develop towards Na enrichment and not depletion, as observed (Tuttle & Bowen, 1958). A shift of the minima horizontally towards the SiO2-Or tieline is caused by H₂O activities of less than 1 (Johannes and Holz, 1990). Consequently, it is conceivable that such a shift is documented by normative pumice data, especially by data from group B. The tieline Qz-feldspars needs to lie on or to the left of the minimum to satisfy a K-enrichment trend of feldspars and not to the right, as observed in group B (Fig. III-7). An increase in water in more evolved compositions is consistent with roofward decrease of percentage of crystals (Table III-2) and by a possible transition from hypersolvus (single feldspar) to subsolvus (two feldspar) crystallization from composition B to the most evolved rhyolite, composition A. Roofward decrease in phenocryst contents can be caused by depression of phase boundaries by higher dissolved volatile contents (e.g., Lipman, 1971, Hildreth, 1979, Ritchie, 1980, Mahood, 1981). Extremely sparse, but nicely euhedral oligoclase (Ab₇₀) microphenocrysts are found in group A which would correspond to Ab₃₀Or₇₀ sanidines (Table III-3) (Fuhrman & Lindsley, 1988). Although such sanidines were not found, they would be the continuation of the evolutionary trend of the hypersolvus feldspars.

Oxygen fugacity

Oxygen fugacities can be calculated using the assemblage quartz-magnetitefayalite (Carmichael et al., 1974, p.109). Since temperature and pressure are crudely known, only a relative distance between the two fayalite bearing rhyolite compositions is relevant. The oxygen fugacities for composition E (#RT173C) is 0.12 log units below composition D (#RT173B). No fayalite, only magnetite and quartz are observed in composition C and B indicating that evolution path of high-silica rhyolites crossed the FMQ equilibrium. More oxidizing conditions can be induced at constant temperature and melt composition with an increase of oxygen fugacity or at constant oxygen fugacity by a temperature decrease (Wones & Eugster, 1965). Despite the hindering of placing a precise path in T vs. logfO₂ space for the different high-silica rhyolite compositions, progressive more oxidizing conditions with higher degree of evolution (E to D to C & B rhyolites) can be documented.

Progressive increase in Mg observed in the clinopyroxenes are opposite to differentiation trends commonly observed in comagmatic suites (viz parallel to FMQ e.g., Carmichael, 1967, Deer et al., 1978, p. 330-342, Carbonin et al., 1991) and are consistent with increasing relative degree of oxidation with evolution in Rattlesnake Tuff. Crystallization under progressively more oxidizing conditions with differentiation relative to delta T (Wones & Eugster, 1965, fig. 13) could produce the observed decrease of Fe/Mg ratios of the clinopyroxenes. Other ferromagnesia silicates, except biotites from group B (#RT34E), also indicate decrease in their Fe/Mg ratio, although not as pronounced as for the clinopyroxenes. Consequently, the decrease of the

ulvöspinel component in the magnetites with evolution of the magmas is possibly not governed by lower temperature but by increasingly oxidizing conditions.

Partition coefficients

Partition coefficients (K) determined from mineral/glass pairs are in table III-6 and are displayed in figure III-8. All listed concentrations for mineral separates of feldspars, clinopyroxene and magnetites are uncorrected for any inclusions (Table III-4). Where inclusions seem to have affected the values, it is discussed below. An upperlimit value (Table III-4) is reported for some trace elements which have values below detection. Upper concentrations limits are useful for evaluation of inclusions and, sometimes, for establishing maximum partition coefficients.

Feldspar

Partition coefficients for anorthoclases and Na-sanidine separates agree well with published partition coefficients for other alkali feldspars (Leeman & Phelps, 1981, Mahood & Hildreth, 1983, Nash & Crecraft, 1985, Stix & Gorton, 1990). REE patterns have a negative slope until Sm and are flat between Sm to Lu with a strong positive Eu-anomaly; coefficients are above 1 for Eu and range from 0.17 to 0.01 for the rest of the rare earth elements. Eu partition coefficients progressively decrease from least evolved high-silica rhyolite-E to group-B composition. K_{Sr} are as high as 13.5. XRF whole rock data were used to calculate Sr partition coefficients in cases where Sr could be determined in the feldspar separate (Table III-4). KBa increases progressively from 7.0 to 19.6 with differentiation. These high coefficients for Ba are real because the processes which could generate high Ba partition coefficients, as outlined by Mahood & Stimac (1990) can be excluded. Phenocryst contents are very low (Table III-2), feldspar crystals are unzoned, alkali feldspar is the ubiquitous Ba carrier, and all feldspars come from a single zoned magma chamber. Ba concentrations in the feldspar are below values where deviation from Henry's Law might introduce variability (Long, 1978).

Fig. III-7 Qz-Ab-Or ternary

Normative pumice compositions and corresponding feldspars. Scatter within each group is likely due to post-depositional ion-exchange. Crosses indicate minima for 0.5, 1, 3, 5 kb pressure at water saturated conditions from Tuttle & Brown (1958). Vector indicates direction of shift of minima under water-undersaturated conditions after Johannes & Holtz (1990). Capital letters indicate high-silica rhyolite group.



Table III-4 INAA bulk mineral and glass analyses

Given error is the analytical error and is a measure for the "distinctness" of the analyzed peak. < : maximum possible concentration for the element which was undetected at a confidence level of three sigma; therefore it is an upper limit concentration value, actual concentration may be less. n.d. : not detected, given instead upper limit. Bromine values are reported to indicate removal of almost all Br-containing heavy liquid residues from mineral surfaces. Feldspar samples consisted of about 70-100 feldspar grains in samples #RT173C, -173B, and -165A and ca. 40 grains in sample #RT34E. Analyzed clinopyroxene separates consisted of about 100 to 300 grains. Each magnetite separate consisted of more than 150 grains.

Table III-4:

	bulk glass								
sample	RT173H	RT34E	RT165A	RT173B	RT173C				
HSRgrp	A	В	С	D	E				
		oxides in	weight %, Mn	O in ppm					
TiO2	0.17 ± 0.03	0.12 ± 0.04	0.14±0.03	0.17±0.04	0.14±0.04				
A12O3	12.54 ± 0.05	12.46±0.05	12.08 ± 0.05	12.23 ± 0.05	12.42±0.05				
FeO*	0.70±0.01	1.22 ± 0.01	1.31±0.01	1.65 ± 0.01	1.77 ± 0.01				
MnO	800±2	780±2	790±2	850±3	860±3				
MgO	n.d.	n.d.	0.38±0.04	0.06±0.006	0.05±0.005				
CaO	n.d.	0.40±0.05	0.50±0.05	0.43±0.05	0.62±0.06				
Na2O	3.56±0.01	3.21±0.01	3.74±0.01	3.88±0.01	4.30±0.01				
K20	4.8±0.1	5.2±0.1	4.8±0.1	4.3±0.1	3.9±0.1				
		trace	elements in	ppm					
Cs	4.25±0.06	3.24±0.05	2.75±0.06	2.35±0.05	2.37±0.05				
Rb	115±1.6	89±2	74±2	62±2	66±2				
Th	9.01±0.05	7.53±0.05	6.78±0.05	5.97±0.05	5.74±0.05				
U	4.57±0.11	3.5±0.16	3.00±0.13	2.79±0.16	2.44±0.15				
Sr	n.d.	n.d.	n.d.	n.d.	n.d.				
Ba	49±8	167±23	357±21	1120 ± 29	1640±28				
La	19.7±0.1	38.3±0.2	48.2±0.14	57.2±0.2	53.7±0.2				
Ce	52±0.4	97.8±0.5	113±0.5	119±0.4	123.0 ± 0.5				
Nd	29±1	51±2	58±3	63±2	59±2				
Sm	9.16±0.01	13.40±0.01	14.00 ± 0.01	14.30±0.01	13.80±0.01				
Eu	0.62 ± 0.01	1.22±0.02	1.34 ± 0.02	1.93 ± 0.02	2.17 ± 0.02				
Tb	2.25±0.02	2.71±0.02	2.42 ± 0.02	2.24 ± 0.02	2.20 ± 0.02				
Dy	15.0±0.4	15.8±0.4	14.9±0.3	13.3±0.3	12.5±0.3				
Yb	10.10±0.05	9.53±0.06	8.5±0.05	7.93±0.06	7.75±0.07				
Lu	1.54 ± 0.01	1.45±0.01	1.28 ± 0.01	1.23 ± 0.01	1.19 ± 0.01				
Sc	3.75±0.01	3.60±0.01	3.32±0.01	3.44 ± 0.01	3.73 ± 0.01				
Zr	170±17	310±16	304±7	389±16	424±16				
Hf	6.63±0.05	9.31±0.06	9.55±0.06	10.40±0.05	10.40±0.05				
Та	2.05±0.03	1.75±0.03	1.46±0.03	1.3±0.02	1.32 ± 0.02				
Cr	4.9±0.5	4.0±0.5	4.4±0.5	4.59±0.5	6.3±0.5				
Co	0.12±0.02	0.45±0.02	0.18±0.02	0.12±0.01	0.16±0.02				
Zn	87±8	108±7	97.3±7	108±6	101±6				
W	1.9±0.3	1.4 ± 0.4	2.0±0.3	2.4±0.3	1.3±0.3				
As	4.6±0.3	4.3±0.4	4.5±0.3	4.0±0.3	3.9±0.3				
Sb	1.48±0.05	1.38±0.05	1.31±0.05	1.25 ± 0.05	1.17 ± 0.04				
Cl	483±72	586±78	1103±73	512±82	1246±92				
Br	3.0±0.3	3.1±0.3	5.4±0.3	3.7±0.3	4.0±0.3				

Table III-4, Continued:

		bulk f	feldspar	
sample	RT34E	RT165A	RT173B	RT173C
HSRgrp	В	С	D	E
	oxide	s in weight	%, MnO in	ppm
TiO2	n.d.	n.d.	n.d.	n.d.
A1203	20.0±1.6	20.19±0.08	21.30±0.09	21.17±0.11
FeO*	0.24±0.005	0.24±0.003	0.23 ± 0.004	0.23±0.003
MnO	10 ± 1	18±0.5	6.5±0.7	8.7±0.8
MgO	0.14±0.02	n.d.	n.d.	n.d.
CaO	n.d.	n.d.	n.d.	1.09 ± 0.12
Na2O	5.82±0.01	6.36±0.01	8.12±0.01	8.76±0.01
K20	7.9±0.2	7.5±0.1	5.1±0.1	4.1 ± 0.1
		trace eleme	nts in ppm	
Cs	0.13±0.04	0.07±0.02	0.14 ± 0.04	0.20 ± 0.04
Rb	30±1	24.8±0.9	12.3±0.8	8.3±0.9
Th	< 0.3	0.20±0.02	n.d.	n.d.
U	< 1.9	< 1.3	n.d.	n.d.
Sr	n.d.	n.d.	162±13	246±16
Ba	2580±41	5680±34	11000±44	12300±37
La	6.5±0.1	4.86±0.08	3.98±0.09	4.00±0.06
Ce	10.2±0.2	7.7±0.2	4.4±0.2	3.9±0.2
Nd	n.d.	n.d.	n.d.	n.d.
Sm	0.11±0.01	0.265 ± 0.007	0.22±0.01	0.12 ± 0.008
Eu	4.58±0.04	6.22±0.03	9.60±0.04	12.20±0.05
Tb	0.066±0.019	0.052±0.010	0.061±0.015	< 0.14
Dy	n.d.	n.d.	n.d.	n.d.
Yb	0.18±0.05	0.22±0.06	0.18±0.03	0.09±0.02
Lu	0.018 ± 0.004	0.030 ± 0.006	< 0.045	0.016±0.005
Sc	0.058±0.003	0.098 ± 0.002	0.049±0.003	0.051±0.003
Zr	n.d.	n.d.	n.d.	n.d.
Hf	0.27±0.03	0.40 ± 0.02	0.48±0.03	0.29±0.06
Ta	0.11 ± 0.04	0.05±0.01	0.06 ± 0.02	0.06 ± 0.02
Cr	9.4±0.4	2.1±0.3	5.9±0.4	7.0±0.5
Co	0.39±0.03	0.09 ± 0.02	0.17±0.02	0.12±0.02
Zn	< 15	< 10	< 7.5	< 13
W	3.3±0.6	1.9 ± 0.6	5.5±0.6	< 3.6
As	< 4.2	< 3.3	< 4.2	< 3.6
Sb	0.31±0.03	0.11 ± 0.01	0.11±0.02	0.15 ± 0.02
CI	n.d.	n.d.	n.d.	n.d.
Br	(5.7±0.5)	(2.6±0.6)	(3.6 ± 1.0)	(2.6 ± 0.4)

sample	RT34E	RT165A	RT173B	RT173C
HSRgrp	В	С	D	E
	oxides	in weight	%, Na2O in	ppm
TiO2	n.d.	n.d.	n.d.	n.d.
A12O3	0.73 ± 0.08	0.85±0.09	0.89 ± 0.08	0.92±0.08
FeO*	21.30±0.06	24.50±0.07	26.40±0.05	28.80±0.09
MnO	2.99±0.01	2.70±0.01	2.45±0.01	2.21 ± 0.01
MgO	8.0±1.2	5.5±1.1	2.3±0.6	2.3±0.4
CaO	18.8±0.8	17.2±0.7	19.2±0.5	18.4±0.7
Na2O	6100±20	6000±20	5000±20	4800±20
K20	n.d.	n.d.	n.d.	n.d.
		trace elem	ents in ppm	
Cs	n.d.	n.d.	n.d.	n.d.
Rb	n.d.	n.d.	n.d.	n.d.
Th	< 3	17.8 ± 3	< 2	< 3.3
U	< 14	< 8	< 6.6	< 8.4
Sr	n.d.	n.d.	n.d.	n.d.
Ba	n.d.	n.d.	n.d.	n.d.
La	36.4±0.3	534±0.5	51.5±0.2	48.0±0.2
Ce	152 ± 4	1210 ± 4	186±2	190±3
Nd	215±9	667±12	227±5	192±7
Sm	81.6±0.1	146±0.2	74.6±0.1	66.2 ± 0.1
Eu	6.08±0.09	10.4±0.1	7.38±0.08	7.78±0.12
Tb	13.2 ± 0.2	15.6 ± 0.2	8.72±0.09	8.41±0.14
Dy	99±4	97±4	69±3	63±5
Yb	47.1±0.6	52.2±0.4	37.4±0.4	36.8±0.3
Lu	8.78±0.08	9.07±0.05	6.99±0.06	6.43±0.06
Sc	496±0.01	467±0.01	385±0.01	361±0.01
Zr	474±120	1300±160	5860±120	810±120
Hf	8.8±0.5	34.3±0.6	19.6±0.4	15±0.5
Та	4.9±0.2	0.82±0.14	0.38±0.08	0.60±0.19
Cr	58±8	47±7	19.2±6	42±7
Co	4.5±0.3	6.6±0.3	3.1±0.2	4.3±0.2
V	n.d	n.d.	n.d	n.d.
Zn	1240±90	1170±87	994±52	1000±81
W	1170±2	23±2	12±1	92±2
As	< 4.8	< 4.8	< 6.6	< 8.1
Sb	< 1.8	< 3.3	< 1.6	< 1.6
Br	(2.4±0.6)	(4.7±1)	(4.8±0.6)	(4.8±0.6)

Table III-4, Continued:

	bulk magnetite							
sample	RT34E	RT165A	RT1733	RT173C				
HSRgrp	В	С	D	E				
	oxides	in weight	%, Na2O in	ppm				
TiO2	11.0±0.3	14.1±0.2	17.0±0.4	19.4±0.2				
A12O3	0.66±0.06	0.65±0.05	0.87±0.09	0.84±0.04				
FeO*	80.4±0.1	78±0.1	76.5±0.2	72.2±0.1				
MnO	2.02±0.01	1.89±0.01	1.74 ± 0.01	1.59 ± 0.01				
MgO	n.d.	n.d.	n.d.	n.d.				
CaO	< 0.19	< 0.21	< 0.29	< 0.05				
Na2O	62±1	155±2	177±2	97±1				
K2O	n.d.	n.d.	n.d.	n.d.				
		trace elem	ents in ppm					
Cs	n.d.	n.d.	n.d.	n.d.				
Rb	n.d.	n.d.	n.d.	n.d.				
Th	0.62 ± 0.17	2.7±0.2	0.74±0.15	1.46 ± 0.13				
U	< 1.8	2.9±0.2	< 2.4	1.78±0.12				
Sr	n.d.	n.d.	n.d.	n.d.				
Ba	n.d.	n.d.	n.d.	n.d.				
La	8.49±0.06	47±0.1	19.9±0.1	27.7±0.1				
Ce	19.3±0.7	116±1	51.4±0.9	69.1±0.7				
Nd	7.9±1.6	79±3	37±7	62±2				
Sm	1.52 ± 0.01	20.50 ± 0.02	9.31±0.02	14.00 ± 0.01				
Eu	0.157±0.02	1.33±0.03	0.88±0.04	1.48 ± 0.03				
Tb	0.23±0.05	2.66±0.08	1.33±0.09	1.94±0.06				
Dy	n.d.	16±3	n.d.	n.d.				
Yb	0.88±0.08	10.9±0.1	2.29±0.09	8.62±0.08				
Lu	0.15±0.02	1.81±0.02	0.30±0.02	1.47±0.02				
Sc	23.30±0.05	24.50±0.05	25.70±0.05	26.80±0.03				
Zr	272±65	2570±80	446±71	3210±22				
Hf	6.98±0.17	77.2±0.2	13.7±0.3	93.8±2				
Та	2.45±0.09	3.12±0.09	4.68±0.13	4.90±0.07				
Cr	19±2	23±2	23±2	20±2				
Co	5.8±0.1	5.57±0.09	6.01±0.13	5.34±0.07				
v	76±6	78±6	73±9	80±4				
Zn	3520±32	3180±29	3130±38	3150±22				
W	< 1.6	1.3+0.2	1.3+0.2	0.79 ± 0.16				
As	< 0.9	< 1.4	< 1.8	< 1.1				
Sb	0.36±0.09	0.28±0.07	0.73±0.1	0.25 ± 0.06				
Br	(2.2 ± 0.3)	(2.1 ± 0.3)	(3.1 ± 0.3)	(0.8±0.2)				

Table III-5 Mineral trace element contributions

Trace element contributions from minerals to whole rock pumice analyses calculated from modal proportions with trace element concentrations of bulk mineral analyses in parts per billion (ppb). Nil: for contributions less than 1 ppb. Values in parentheses are contributions where trace elements from zircon are added to budget. Zircon contributions were calculated assuming it occurs only as inclusions in mafic phases at 1 wt% of the mafic phases (see text). Hf and Zr concentrations in zircons are from Rattlesnake Tuff zircons; concentrations of other elements in zircon obtained with partition coefficients from Mahood & Hildreth (1983).

Elemental mineral contribution								
sample	34E	165A	173B	173C				
HSRgrp	В	С	D	E				
		parts per	billion					
Cs	nil	1	1	1				
Rb	8	183	108	43				
Th	nil (4)	7(11)	nil (6)	1(6)				
U	nil (7)	1(7)	nil (10)	1(11)				
Sr			1416	1284				
Ba	645	41805	96140	64206				
La	6(8)	198(199)	71(75)	62(66)				
Ce	18(24)	426(432)	162(174)	166(180)				
Nd	17(18)	207(209)	140(143)	144(147)				
Sm	6(7)	48(49)	47(48)	47(49)				
Eu	2(2)	49(49)	88(89)	69(70)				
Tb	1(2)	5(6)	6(7)	6(7)				
Yb	4(33)	19(45)	23(66)	27(75)				
Lu	1(6)	3(8)	4(12)	5(13)				
Sc	41	138	223	232				
Zr	107(2724)	1032(3678)	505(5250)	1874(7353)				
Hf	3(64)	33(82)	21(102)	51(135)				
Та	1	1	3	3				
Sb	nil	1	1	1				
Zn	1037	1154	1846	2054				
Mn	5790	9500	16485	18135				
Cr	12	34	72	71				
Co	2	4	6	6				

				Partition	coefficien	its from	mineral/gl	ass pairs				
	feldspar			clinopyroxene			magnetite					
sample	RT34E	RT165A	RT173B	RT173C	RT34E	RT165A	RT173B	RT173C	RT34E	RT165A	RT173B	RT173C
HSRgrp	B	С	D	E	В	С	D	E	B	С	D	E
Cs	0.04	0.03	0.06	0.08								
Rb	0.34	0.34	0.20	0.12								
Th		0.03				2.63			0.08	0.39	0.12	0.25
U										0.97		0.73
Sr			13.50	10.70								
Ba	19.55*	15.91	9.82	6.95*								
La	0.17	0.10	0.07	0.07	0.95	11.08	0.90	0.89	0.22	0.98	0.35	0.52
Ce	0.10	0.07	0.04	0.03	1.55	10.71	1.56	1.54	0.20	1.03	0.43	0.56
Nd					4.22	11.50	3.63	3.27	0.16	1.37	0.60	1.05
Sm	0.01	0.02	0.02	0.01	6.09	10.43	5.22	4.80	0.11	1.46	0.65	1.01
Eu	3.8	4.6	5.0	5.6	5.0	7.8	3.8	3.6	0.1	1.0	0.5	0.7
Tb	0.02	0.02	0.03		4.87	6.45	3.89	3.82	0.08	1.10	0.59	0.88
Dy					6.24	6.54	5.23	5.04		1.09		
Yb	0.02	0.03	0.02	0.01	4.94	6.14	4.72	4.75	0.09	1.28	0.29	1.11
Lu	0.01	0.02		0.01	6.06	7.09	5.68	5.40	0.10	1.41	0.24	1.24
Sc	0.02	0.03	0.01	0.01	137.8	140.7	111.9	96.8	6.47	7.38	7.47	7.18
Zr					1.53	4.28	1.51	1.91	0.88	8.45	1.15	7.57 .
Hf	0.03	0.04	0.05	0.03	0.95	3.59	1.88	1.44	0.75	8.08	1.32	9.02
Та	0.06	0.03	0.04	0.05	2.82	0.56	0.29	0.46	1.40	2.14	3.60	3.71
Sb	0.23	0.08	0.09	0.13					0.26	0.21	0.58	0.21
Zn					11.48	12.02	9.20	9.90	32.59	32.68	28.98	31.19
Mn	0.01	0.02	0.01	0.01	39.36	33.04	29.18	25.00	25.26	23.54	20.12	18.49
Cr	2.34	0.47	1.29	1.11	14.34	10.54	4.18	6.67	4.79	5.09	4.90	3.16
Co	0.86	0.48	1.40	0.74	9.96	36.39	25.67	26.63	12.87	30.94	50.08	33.38

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* Ba of glass of # 34E and 173C is whole rock data; for 173C corrected for minerals

Mineral partition coefficients calculated from mineral/melt pairs. Table III-6 Partition coefficients





Coefficients are between 0.12 to 0.34 for Rb and 0.03 to 0.08 for Cs; those for Rb increase slightly towards more evolved compositions. Sc, Mn, and Hf partitions similarly to reported values for comparable rhyolites (Mahood & Hildreth, 1983) with value for K_{Sc} and K_{Mn} ranging from 0.01 to 0.03 and for K_{Hf} from 0.03 to 0.05; Ta partition coefficients are slightly higher. Values for Sb are among the first to be reported. Partition coefficients for Cr and Co are significantly higher compared to other transition metals (Fig. III-8). The internal scatter among Cr and Co could be within analytical uncertainty, however the overall jump to partition coefficients close to 1 seems real, especially in the light of low Mn and Sc values. Any mineral inclusions increasing Cr and Co concentrations would have increased Mn and Sc as well. Co partition coefficients for feldspars, tens times or more compared to Sc and Mn, have also been reported by Mahood & Hildreth (1983).

Clinoyroxene

Partition coefficients of Sc increase progressively with evolution of the magma; the range around 100 is similar to other Fe-rich clinopyroxenes of high-silica rhyolites (Hildreth & Mahood, 1983). A progressive increase of K with differentiation is also evident for Mn (Table III-6, Fig. III-5, III-8). Despite some scatter, a minor increase is seen in Cr, Zn, and possibly Ta partitioning. The jump of Ta partition coefficients from values around 0.5 to 2.82 in pumice group B might be real because no suitable trace phase or contaminating liquid seems appropriate to selectively enrich Ta, however it is taken with caution.

Partition coefficients for REE elements in three out of four clinopyroxene separates from the Rattlesnake Tuff show patterns similar to augites from rhyodacite and to hedenbergites (Nash & Crecraft, 1985, Mahood & Stimac, 1990). The magnitude of partition coefficients for LREE and MREE compares well with partition coefficients determined for individual spots by ionprobe for similar clinopyroxene (Sisson, 1991) so the partition coefficients are not governed by inclusions of REEbearing trace phases (c.f. Michael, 1988). One sample (#RT165A), however, is unusually enriched in LREE, but has values for the MREE and HREE only slightly above the values for other samples. The LREE enrichment in sample RT165A is most likely the result of undetected minute inclusions of chevkinite. HREE patterns of clinopyroxenes are flat within analytical uncertainties and do not indicate a downwarp of the pattern as observed in other clinopyroxenes from high-silica rhyolites (e.g. Sisson, 1991). Zircon inclusions are not a considerable contributor of HREE because partition coefficients for Zr and Hf (uncorrected for any inclusion) are below the ones for HREE for all samples. Calculating the potential contribution from zircon inclusions by assigning all Hf to zircon in the clinopyroxenes from sample RT34E, which has HREE concentrations (or K's for HREE) comparable to the other separates but also has lowest Hf content (8 ppm) would reduce the contribution of HREE from zircon to 8.6 % (Hf_{zircon} = 10000 ppm; K_{Lu} in zircon 650 from Mahood & Hildreth, 1983). The clinopyroxenes with the lowest Hf at equal HREE gives the lowest maximum value of a potential HREE contribution from zircon inclusions. Therefore a flat HREE pattern is interpreted to reflect intrinsic clinopyroxene REE partitioning. On the other hand, only the lowest Zr and Hf values are likely to be close to true Hf and Zr partitioning into clinopyroxene (Mahood & Stimac, 1990; Sisson, 1991).

Magnetite

Magnetites are enriched in the transition metals Zn, Mn, Sc, Cr, Co, and Ta. Zn, Mn, and Co have similar partition coefficients, around 20 to 30, with higher values for Zn than for Mn. Zn partition coefficients are nearly constant while coefficients for Mn indicate a slight progressive increase with higher degree of evolution of the pumice composition (Table III-6, Fig. III-8). The scatter of Co values (mainly caused by high Co in glass analyses of sample RT34E and slightly lower Co in #RT173B) is likely to be due to the uncertainty of the low Co concentration in the glass. Partition coefficients for Cr and Sc are 4 and 7, respectively and are nearly constant through the compositional range of rhyolite compositions. Ta partition coefficients decrease from 3.7 to 1.4 towards more evolved liquids. In general, partition coefficient and interelement variations in partition behavior for Sc, Cr, Co, Zn, Mn, and Ta are comparable to other felsic systems (Mahood and Hildreth, 1983, Woerner et al., 1983, Mahood and Stimac, 1990). Sb is incompatible in magnetite.

The range of partition coefficients for REE, U, Th, Zr, and Hf of magnetite separates is most likely variably affected by trace phase inclusions which easily escape detection in opaque magnetites. However, the extent of the contribution of REE elements to obtained concentrations in bulk magnetite samples must have been small. Assigning all Hf or Th to zircon inclusions would overcorrect Lu values to values below zero. Maximum values for undetectable Ca in the magnetite samples (Table III-4) reduces, for instance, Sm contribution from apatite in the bulk magnetite sample

constant as K_{Yb} and K_{Lu} . REE elements are thought to reside mostly in the M2 site in pyroxenes where they substitute for Ca (Gallahan and Nielsen, 1992). Ca and Na, also hosted in the M2 site, are nearly constant in the range of clinopyroxenes. A correlation seems to exist between REE partitioning and Al and Ti (Gallahan and Nielsen, 1992). Al in clinopyoxenes from group E to B are almost constant (Table III-3, III-4). Ti concentrations in clinopyroxene decrease generally from ca. 0.25 weight % in E to 0.15 in B and Ti in liquid decreases from 0.17 weight % in E to 0.11 in B and A. Therefore K_{Ti} decreases subtly from 1.5 to 1.4, that is, opposite to the small variation in K_{Sm} .

On the other hand, a progressive increase in partition coefficients from group E to B is noticeable for Sc and Mn. Cr, Zn, and Co indicate similar trends but analytical uncertainties of these elements in the glass or in the pyroxenes reduce the observed variations in partition behavior. The increase in K_{Sc} and K_{Mn} in cpx closely correlates with changing mineral chemistry of the clinopyroxenes which is indicated by nearly constant compound partition coefficients of Sc/Fe* and Mn/Fe* (Fig. III-9) and is consistent with partitioning of Sc and Mn into the M1 site (Carbonin et al., 1991, Gallahan & Nielsen, 1992). It is likely that the stronger partitioning of Sc and Mn is a response to a decrease of $Fe^{2+/}Fe^{3+}$ in the melt making incorporation of Sc and Mn over Fe^{3+} favorable. Mn typically increases with Fe content of the pyroxenes (Deer et al., 1978, p. 325) but opposite trends are indicated for the Rattlesnake Tuff clinopyroxenes (Fig. III-5).

In magnetites, K_{Sc} , K_{Zn} , K_{Cr} and concentrations of Co are constant within analytical uncertainty. K_{Ta} decreases progressively with degree of evolution. Zn, Mn, Cr, and Co can fairly readily substitute for Fe whereas Ta⁵⁺ ion with its high charge is likely to primarily substitute for Ti. Ta partitioning is best explained by decreasing Ti content of the magnetites as supported by nearly constant compound partition coefficient Ta/Ti (Fig. III-9). K_{Mn} increases in magnetites.

Mn partitioning trends in magnetites are similar to trends in pyroxenes, fayalites, and biotites indicating Mn enrichment with degree of evolution of high-silica rhyolite magma. The only exception is biotites from sample RT34E. Possible reasons for these "reverse trends" (increased Mn concentration at decreasing Fe*) in Rattlesnake Tuff mafic silicates, especially clinopyroxenes, include increase of melt polymerization, less volatile complexing, and lower temperature. Evidence against these reasons is found in internal inconsistencies of partition coefficients in magnetites. Higher partition coefficients of Mn should correspond with higher partition coefficients of Zn, Sc, Cr, and Co which is not observed. Another possible reason is the existence of significant Mn³⁺ which could be preferentially taken up compared to divalent ions, such as Zn²⁺,

although Mn is thought to exist completely as Mn²⁺ ion in most terrestrial magmas (Carmichael & Ghiorso, 1990, Fig. 5). Observed partitioning of manganese into silicates and oxides from the Rattlesnake Tuff leads to Mn concentrations among the highest reported for either Ca-rich pyroxenes, Fe-rich olivines, or Ti-magnetites (Carmichael, 1960, 1967, Deer et al., 1978, 1982, Mahood, 1981). In conclusion, changes in partition behavior in magnetites, clinopyroxenes, and feldspars correlate closely with crystal chemical changes.

Chemical changes

The existence of progressively more evolved high-silica rhyolite liquids allows to investigate of how single elements behaved through the course of differentiation. To this purpose, enrichment of all the different high-silica rhyolite compositions relative to the least evolved rhyolite group E are plotted in figure III-3. Changes in the rate of enrichment from one high-silica rhyolite composition to the next more evolved composition for selected elements are illustrated by comparing enrichment between high-silica rhyolite compositions (Fig. III-10).

Ba enrichment rates decrease first, but then stay fairly constant (Fig. III-10). Eu enrichment rates are constant during early stages of differentiation but the Eu enrichment rate is partially decoupled from the Eu anomaly in the most evolved composition where the rate is comparable to La. Rates indicating development or retreat of an Eu-anomaly (Eu/Eu*) are almost constant throughout. The ratios of enrichment rates, rate_{Eu}/rate_{Ba}, are close to the ratios of the partition coefficients K_{Ba}/K_{Eu} for feldspars of the less evolved composition (normalizing compositions in figure III-10) for the first three intervals. Therefore for the first three intervals, removal of observed feldspars alone without any additional phase could explain the enrichment pattern of Ba and Eu.

La enrichment rates during the first two increments are close to one (constant concentration) but decrease with continued differentiation. Zr enrichment rates are also constant through the same intervals as La and decrease during the last two increments. The decrease in enrichment rates for La and Zr indicates that a LREE-enriched phase and zircon, or other phases enriched in these elements, become more important as evolution proceeded, assuming that the fractionation amount between adjacent compositions did not change. More fractionation could decrease enrichment rates for elements with a bulk partition coefficient greater than one.

The overall change of enrichment rates for La and Zr has a mirror image in the rates of enrichment for the incompatible elements, exemplified by Cs, Rb, U, Th, and Ta. Enrichment rates for these elements are fairly constant, sometimes close to analytical uncertainties until reaching the last evolutionary step in which enrichment rates increase (Fig. III-10) accounting for about half of the maximum enrichment for these elements (Fig. III-3). The increase in enrichment rates for the incompatible elements lags one step behind the enrichment rate changes in La and Zr. In case of the incompatible elements, an increase in enrichment rates could be ascribed to a change in the mass proportions where the fractionation amount is constant but increases considerably during the last evolutionary step. This interpretation assumes a derivation of the liquids through fractionation, that only liquid-crystal partitioning is important, and that no assimilation occurred (see below). A change in the mineral assemblages to explain an increased incompatibility is unlikely because calculated bulk partition coefficient (D) are virtually zero for Cs (Table III-7). Similarly, reduction of D for Rb, for U, and for Ta is unlikely because of evidence for continued feldspar removal (main contributor to D_{Rb}), zircon fractionation (U host), and increasing D's for Ta with differentiation (Table III-7).

Fig. III-9 Compound partition coefficients



Fig. III-10 Enrichment rates for selected elements

Rates were determined by calculating enrichment factors between group E and D (D/E), group D and C (C/D), group C and B (B/C), and between group B and A (A/B). Chemical averages were used except, group-C is represented by a single sample, RT165A, due to larger scatter within group C.



Discussion

Differentiation through some kind of fractionation process along a continuous liquid line of descent is the most reasonable process to generate a series of progressively more evolved liquids coexisting in one chamber. Other fractionation scenarios include: (a) all more evolved rhyolite liquids are directly derived from one single parent, the least evolved group-E composition, or (b) all high-silica rhyolites, including composition-E, fractionated from some rhyolite composition which did erupt. A number of models to derive the variation of high-silica rhyolites are discussed below.

Constraints

Eruption of different high-silica rhyolite magmas was short enough to generate a single cooling unit. The minor precursory fallout and many basal and distal tuff sections which are composed entirely of white (clear) shards and white pumices indicate that the first magma to come out of the conduits was the most evolved. On the other hand, pumices representing the whole compositional rhyolite spectrum were sometimes collected at the same tuff section commonly 10 to 20 m thick (e.g. sample location "173") indicating that all observed rhyolite magmas coexisted in some geometric arrangement in one chamber. The eruption sequence can be interpreted to reflect, generally, the inverted stratification of the magma column (Smith, 1979). The high-silica rhyolites were underlain by dacites and more primitive magma discussed elsewhere (Chapter II).

Rattlesnake Tuff pumice data cluster in 4 to 5 chemical groups indicating that several distinct magma compositions existed. Although rheological differences among magmas can introduce compositional gaps during eruption from zoned chambers (Blake, 1981, Blake & Ivey, 1986, Spera et al., 1986), the pumice clusters reflect intrinsic zonation of the magma column and are not an artifact of eruption dynamics. Among the factors governing magma rheology, only volatile gradients are not known. On the other hand, all high-silica rhyolite compositions are very similar in major element composition and crystal content; only strong volatile gradients might have introduced rheological differences. Therefore, it is unlikely that strong rheological differences among high-silica rhyolite compositions existed.

Table III-7 Bulk mineral partition coefficients

D's used for crystal fractionation models in figure III-11.

bulk	partition		coel	ficients,	D	_	
sample	34E	165	5A	173B	173C		
HSRgrp	В	C	2	D	E		
Cs	0.01	0.0)1	0.05	0.07		
Rb	0.09	0.1	9	0.17	0.10		
Ba	5.34	8.8	33	8.65	5.70		
Eu	1.43	2.6	56	4.60	4.97		
Sc	12.20	3.1	0	6.57	9.85		
Та	0.48	0.0)7	0.20	0.33		
Zn	10.60	0.9	0	1.70	3.06		
Mn	10.48	1.1	16	2.46	3.67		

Without fayalite and without trace phases. K_{Eu} for all magnetites are K_{Eu} in 34E since lowest.

 K_{Eu} for cpx of 165A is the one from 173B, because 165Acpx is affected by LREE phase. K_{Ta} for magnetite of 34E exchanged by value for 165A due to uncertainty of high Ta

Fig. III-11 Crystal fractionation model

X-axis shows modeled step, y-axis displays required fractionation amount, cumulative, to explain observed concentration in daughter liquid with observed modal abundances and element concentrations in minerals. Open symbols indicate incompatible elements, solid symbols indicate compatible elements. Selected elements are independent of accessory trace phases. $(D>1) = D_{element} > 1$ and observed daughter concentration is higher than parent concentration therefore concentration in daughter cannot be modeled for this particular interval.


Fig. III-12 Modeling of bulk partition coefficients

Solid dots indicates bulk partition coefficients required to fit daughter compositions of crystal fractionation models to observed compositions of group averages; capital letters indicate high-silica rhyolite groups (Table III-1), amounts of fractionation for each step are based on enrichments of Cs and Rb. Open dots indicate bulk partition coefficients based on best proportions of observed major phases: alkali-feldspar, clinopyroxenes, magnetites, and quartz; best proportions are determined by match with required bulk partition coefficients combined with requirements of major elements. Open triangles indicate bulk partition coefficients with addition of accessory phases (see text). Major element compositions of melts and removed solids with modeled mineral proportions are in table III-8. Partition coefficients for Zr and Hf of magnetite for model D to C are from #RT34E; for model C to B, partition coefficients for REE of clinopyroxenes are those from #RT34E, and K_{Zr} and K_{Hf} are those from #RT34E. Other used partition coefficients are: K's=0.01 for quartz; Ksr=10 of feldspar where not determined; KTh and KU=0.01 for all major phases; KSr and KBa =0.01 for clinopyroxene, magnetites, zircon, and allanite; KREE are interpolated where not available; KZr and KHf =0.3 for clinopyroxene (Sisson, 1991); Ky=KTb for all phases where no Ky available; K_{Nb}=K_{Ta} for all phases; K_{Zn}=K_{Mn} for feldspars; partition coefficients for zircon are values for zircon from Bishop Tuff (Mahood & Hildreth, 1983); apatite partition coefficients were taken from Stimac & Mahood (1990), and K's for allanite are from Bishop Tuff allanite (Mahood & Hildreth, 1983); allanite instead of chevkinite was used as LREE-enriched phase due to a lack of K's for chevkinite. Mn values of group A to E are values determined from glass separates by INAA.

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Fig. III-12:



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cpx=clinopyroxene, mag=magnetite, qu=quartz, zirc=zircon, ap=apatite, alla=allanite. melt=modeled daughter composition, avg X= composition of group averages (Table Table III-8 Fractionation models with non-modal mineral proportions Proportions of accessory phases, zirc, ap, and alla, solely based on trace element III-1), min.%= mineral proportions of removed solid, fsp=alkali-feldspar. frac. %= amount of fractionation, solid = bulk composition of removed minerals, Results of crystal fractionation models with non-modal mineral proportions; total

requirements (see text).

	model E to D			model D to C			model C to B			model B to A		
total frac.	. %	7		13				13			27	
	solid	melt	avg D	solid	melt	avg C	solid	melt	avg B	solid	melt	avg A
SiO2	64 4	75.9	76.0	73.2	76.7	76.9	75.7	77.3	77.0	76.7	77.2	77.4
TiO2	0.67	0.13	0.14	0.98	0.02	0.13	0.65	0.05	0.12	0.561	-0.02	0.11
A1203	20.33	11.79	12.00	13.23	11.86	11.80	11.69	11.85	11.8	11.09	12.08	11.9
FeO*	3.25	1.86	1.68	4.0	1.34	1.48	3.45	1.19	1.22	3.47	0.4	0.77
MnO	0.12	0.09	0.09	0.1	0.09	0.08	0.09	0.08	0.09	0.1	0.09	0.08
MeO	0.09	0.08	0.06	0.01	0.07	0	0.03	0	0.07	0.03	0.08	0.06
CaO	12	0.48	0.57	0.48	0.58	0.37	0.28	0.38	0.33	0.18	0.39	0.28
Na2O	7 37	3.53	3.10	4.94	2.84	3.5	3.78	3.47	3.30	3.38	3.27	3.59
K20	3.99	6.02	6.02	3.06	6.48	5.5	4.28	5.7	5.9	4.54	6.41	5.32
min.%												
fsp	94			65			60			57		
CDX	3			0.4			0.6			0.4		
mag	3			5			4			4		
au	0			30			35.4			38.6		
zire	0.2			0.13			0.2			0.15		
a 10	0			0.04			0.05			0.9		
alla	0.02			0.05			0.13			0.12		

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Fractionation from chamber interior

Classical Rayleigh crystal fractionation of observed phases in modal proportions can be evaluated by modeling trace elements which are not affected by accessory phases such as zircon, apatite, or chevkinite and whose concentration levels are analytically well constrained in minerals and glasses. Elements fulfilling these requirements include Cs, Rb, Ba, Eu, Sc, Zn, and Mn. XRF-data for whole rock pumices, corrected for minerals according to values in table III-5, are partially used for glass concentrations of Rb, Ba, and Zn (Fig. III-11). Europium is included for the first three steps modeled. Despite the high Eu concentration in apatites, the observed relative apatite proportions of <<1% of all crystals reduces its control on Eu-systematics to negligible. KEu for apatite is about 30 (Mahood & Stimac, 1990), only 5-10 times higher than in feldspars and clinopyroxenes. The observed liquid line of descent was modeled successively by deriving each more evolved rhyolite composition from the previous less evolved composition. Composition E was parent to D, D was parent to C, C to B, and composition B was parent to the most evolved composition A. Modeling was done such that for each interval modeled the fractionation amount was adjusted for each element individually to fit calculated daughter concentration to the observed concentration of the selected daughter composition. The used bulk distribution coefficients, table III-7, were calculated with information from table III-2 and III-6. The results are in figure III-11.

If all elements variations could be explained by the same amount of fractionation, the lines in figure III-11 would coincide. Deviation from this indicates that variable fractionation amounts for different elements are required to derive the observed daughter compositions with observed bulk partition coefficients. Parallel lines indicated that the amount of fractionation required deviated in an earlier modeling step, but can be modeled in more evolved compositions by the same fractionation amount. For example, parallel lines are observed for Mn and Zn during modeling rhyolite composition C to B (Fig. III-11). The significance of any deviation between elements in the amount of fractionation required depends mainly on the bulk distribution coefficient, D . If D is close to unity then deviation at fractionation amounts less than 10% are irrelevant, such as deviations between Mn and Zn developed during modeling composition C from composition D. On the other hand, when D deviates considerably from unity, then even a 10% discrepancy is likely to be significant, such as the 10% deviation between Ba and Mn during modeling rhyolite group-D from group E. The main information to extract from figure III-11 can be summarized by three points: (1)

incompatible elements, Cs, Rb, and Ta require similar fractionation amounts (2) discrepancies in fractionation amounts between the transition metals, Sc, Mn, and Zn, to other elements, except Eu, widens towards more evolved liquids with maximum discrepancies of 30 to 50%, and (3) discrepancies exist between Ba and Eu caused by partition coefficients for Eu of clinopyroxenes; K_{Eu} 's in clinopyroxene are great enough to increase D_{Eu} but not D_{Ba} . Crystal fractionation from the chamber interior using observed modal proportions and observed bulk partition coefficients is unlikely because it leads to inconsistencies among trace elements whose distribution is independent of accessory phases.

Boundary layer fractionation

Crystal fractionation in non-modal proportions might occur in boundary layers along the margins of the magma chamber. Because there is no direct evidence for crystal fractionation along walls (or liquid fractionation), liquid fractionation was modeled such as to test whether the results are reasonable and internally consistent therefore providing evidence for validity for such model. The modeling assumed the following: (1) only observed feldspar, clinopyroxene, magnetite, and quartz crystallized as major phases, (2) partition coefficients for major phases are the same along margins as determined (Table III-6), and (3) increases in Rb and Cs concentrations with degree of evolution in Rattlesnake Tuff high-silica rhyolites are completely due to crystal fractionation. The first modeling step included the determination of the amount of Rayleigh fractionation using Cs and Rb variations between each rhyolite composition and their bulk partition coefficients (Table III-7). The amount of fractionation for Rb and Cs varied by 1% for model E to D, 3% for D to C, 8% for C to B, and by 4% for model B to A; intermediate fractionation amounts were further used. During the second step, the fractionation amount from step 1 and the element variations between rhyolite compositions were combined to determine bulk partition coefficients, D for trace elements. The third step involved matching the required bulk partition coefficients from step 2 with calculated D's derived by using K_{Ba} and sometimes K_{Eu} as tracer for alkali-feldspar and K_{Sc} , K_{Mn} , and K_{Zn} for a best combination of clinopyroxene and magnetite. This procedure yielded a certain feldspar, clinopyroxene, and magnetite proportion which was used to calculate the major element composition of a "bulk mineral" subtracted from the parent to yield the major element composition of the daughter. Proportions of quartz were obtained by silica requirement

for the "bulk mineral" to obtain a silica concentration in the daughter compositions within 0.3 weight % SiO₂ of the observed daughter. Accessory phase proportions were determined similarly as in step 3 by using K_{Zr} and K_{Hf} as tracer for zircon, K_{LREE} as tracer of allanite (or generally LREE-enriched phase), and apatite to improve match of D_{MREE} . All described steps were done for the individual intervals composition E to D, D to C, C to B, and composition B to A with modeling results in figure III-13 and table III-8.

The cumulative amount of crystal fractionation from composition E to A is 60% and mineral proportions satisfying requirements for major phases exist with generally good internal consistencies among calculated D's and with majors elements. Compositions of residues are syenitic for modeling rhyolite E to D, and granitic for the others (Table III-8). Proportions of zircon range from 0.13 to 0.2%, allanite from 0.02 to 0.13% and apatite from 0 to 0.9% reproducing well required D_{REE}'s, except D_{HREE}'s for step E to D and D to C, which could be due to using higher K_{HREE} for zircon than in Rattlesnake Tuff zircons. D_{Th,U} and D_{Ta,Nb} are not well reproduced and required D's are commonly lower than modeled for Th and U and commonly higher for Ta and Nb. In conclusion, liquid fractionation of observed phases in non-modal proportions yields a liquid line of descent consistent with gradients in the high-silica rhyolite compositions E through A.

Models with crustal assimilation

Assimilation and contamination with wall rocks have been proposed to explain incompatible element enrichments and isotopic systematics in many ash-flow tuffs (e.g., Woerner et al., 1985, Tegtmeyer & Farmer, 1990, Farmer et al. 1991). Estimated assimilated wallrock for similar size systems as the Rattlesnake Tuff range from 1-18 weight % (Tegtmeyer & Farmer, 1990), 5-30% (Grunder, 1987), to 20-40 weight % (Farmer et al. 1991). Most isotopic variations of single ash-flow sheets result from differences between rhyolite and dacite magmas. The internal isotopic variations of rhyolitic parts of single ash-flow tuffs can be very narrow (Halliday et al., 1984).

If assimilation without fractionation was important in Rattlesnake Tuff, the assimilated material must have had a major element composition similar to rhyolites but higher Rb, Cs, Ta(Nb), Th, U, Y, HREE, stronger Eu/Eu* and much lower Ba, lower Eu, LREE, Zr(Hf), and Sr than composition A to explain element gradients between rhyolite endmembers E and A. The combination of elements required to be enriched or

depleted in the contaminant is internally inconsistent or geologically unreasonable. For example, the rhyolitic assimilant would need to contain much zircon or mafic silicates to enrich magma in HREE and Y; on the other hand, zircon or mafic silicates are the main hosts for Zr and Hf which indicate depletion. Required element concentrations in the contaminant are less than 30 ppm Ba and less than 2 ppm Sr and are geologically unreasonable.

Combined models of assimilation and fractionation (AFC) could explain observed enrichment and depletion trends. Only if the fractionation amount required from fractionation models without assimilation could be reduced when combined with assimilation, AFC-models would be preferred over fractionation models alone. The highest amount of fractionation (55 to 65%) is suggested from enrichments of the elements Cs, Rb, and Ta and depletion of Ba (Fig. III-11), therefore these would be the target elements whose concentrations in the magma need to be modified by assimilation to validate a need for assimilation in modeling. Bulk assimilation (all elements) had to occur in order to enrich mobile elements, Cs and Rb, and immobile elements, Ta and Nb, in the magma solely through assimilation. It is unlikely that element contribution through bulk assimilation would reduce the required fractionation. For example, Ba gradients in Rattlesnake Tuff rhyolites impose high fractionation amounts. An assimilant with 500 ppm Ba would reduce the fractionation amount only by mixing with rhyolite E (1900 ppm Ba) and rhyolite D (1100 ppm) before fractionation would further reduce Ba concentrations. A >500 ppm Ba assimilant is most likely because it corresponds to estimates of Ba values in upper continental crust (Taylor & McLennan, 1985, p.46). Therefore AFC models would reduce the amount of fractionation only for model E to D and D to C. For the more evolved compositions, assimilation of wallrocks with 500 ppm Ba would only increase the amount of fractionation for model C to B and B to A which are the steps requiring ca. 40 of 60% total fractionation based on enrichment of Cs, Rb, Ta(Nb) and depletion of Ba (Fig. III-11). Therefore, assimilation might have occurred but is thought to have played a minor role for the internal elemental variations of the Rattlesnake Tuff high-silica rhyolites.

Progressive partial melting

Accumulation of progressively higher degrees of partial melts is one of the processes which might generate a zoned rhyolite magma. This process is only evaluated with respect to whether it could generate the diversity in high-silica rhyolite magma.

Melting scenarios to produce the least evolved high-silica rhyolite magma (group-E rhyolites) from amphibolites or greywackes are discussed in Chapter II. During progressive melting, the first degree of melt would be represented by the most evolved rhyolite, composition-A, based on higher incompatible element concentrations (e.g. Rb, Th) and higher silica; consequently, the least evolved rhyolite, composition-E, would represent a greater degree of melting. Arguments previously put forward to discard progressive melting scenarios apply in a similar fashion to the Rattlesnake Tuff (c.f. Hildreth, 1979, 1981, Mahood, 1981). Barium concentrations of the high-silica rhyolite composition A is about 30 ppm which is much below the concentration derived from any reasonable source lithology at low melt fraction. Other arguments against progressive melt include the extreme, 60-fold, Ba gradient over a 2.5 weight % silica range from group-E to group-A rhyolite and that the trend for Rb, which is opposite to Ba, indicates only a twofold change. The derivation of the REE pattern of the most evolved rhyolites through low degrees of partial melting would require a source with a flat or depleted LREE pattern, characteristic of tholeiitic basalts, but with MREE and HREE concentrations at about 40 times chondrite which is rarely observed. Also, towards greater degree melts, the REE pattern would change from flat (group A) to LREE enriched (group E) accompanied by a crossover at Tb violating reasonable melting models.

Mixing

Mixing of high-silica rhyolite endmembers could produce a range of high-silica liquids. However, it is inadequate to explain the intermediate high-silica rhyolite compositions (group B to D) of the Rattlesnake Tuff. For example, if rhyolite composition B were a mixture of A and E, required mixing proportions for Ba would be 8% rhyolite-E and 92% group-A rhyolite, on the other hand for La the proportions would be 58% composition E and 42% composition A. Similar discrepancies in mixing proportions exist between other elements and are graphically represented on element-element scatter plots by the pumice clusters defining curvilinear arrays (Fig. III-2). Only macroscopically banded pumices fall near mixing arrays between different high-silica rhyolite compositions (Fig. III-2).

Speculations

From the above discussion of the processes which could generate a compositionally zoned magma chamber as preserved in the high-silica rhyolite pumice clusters of the Rattlesnake Tuff, liquid fractionation (synonymous with boundary layer or convective fractionation) can account best for generating the observed chemical gradients.

Examples of several distinct rhyolitic magma compositions preserved in other ash-flow tuffs have been described from the Grizzly Peak Tuff (Fridrich & Mahood, 1987) and the Toquima Caldera Complex (Boden, 1989). In both examples, suggestions were made that magmas were arranged as density stratified, horizontal homogenous layers with intra-layer convection and stepwise compositional gradients across adjacent layers. One important feature of the invoked intra-layer convection is that is preserves distinct magmas because magmas of different layers only mixed when larger-scale convection occurs (Fridrich & Mahood, 1987). Density stratification of different magmas in the pre-eruptive magma chamber has been inferred for a series of other compositionally zoned ash-flow tuffs (e.g. Hildreth, 1979, Grunder & Mahood, 1988, Wolff et al., 1990). Density stratification seems also to work for the high-silica rhyolite magmas of the Rattlesnake Tuff. Densities progressively decrease with higher differentiation degree. Higher volatile contents towards the top of the chamber, suggested by decreasing mineral content and by calculated enrichment of volatiles through fractionation, would accentuate the density gradient. Therefore, a pre-eruptive scenario with several distinct high-silica rhyolite layers which internally convect, similar to models for the Grizzly Peak Tuff and the Toquima Caldera Complex (Fridrich & Mahood, 1987, Boden, 1989), is envisioned for the Rattlesnake Tuff magma chamber.

Differentiation processes operating within the interior of the magma chamber, whether driven by crystal fractionation or volatile complexing, are inconsistent with the development of a layered cell configuration. Interior processes would generate smooth gradients instead of stepwise gradients or discontinuous trends among different elements as discussed for the Rattlesnake Tuff rhyolites. In addition, the rhyolitic magma column would have needed to develop its full compositional range before breaking down into several compositional layers in order to combine differentiation processes operating vertically (?) in the chamber interior with horizontal convection layers. On the other hand, many arguments speak in favor of differentiation processes operating along the chamber walls (e.g. Tait et al., 1989) An important parameter for differentiation processes occurring along the walls is the shape of the chamber, especially its length-to-width ratio, because it governs where this wall fractionation process might primarily occur. Flat cylindrical shapes for large rhyolitic bodies are deduced from diameters and downdrop distances in calderas and cauldrons (Smith, 1979, Spera, 1981). In such instances, the chamber top is several times the thickness of the rhyolitic magma column. With such a given geometry, liquids derived along the top of the chamber would be several times the mass of liquids produced along the sides if one assumes the same fractionation conditions for sides and chamber tops.

A model in agreement with observations from the Rattlesnake Tuff high-silica rhyolite magmas is proposed (Fig. III-13). A reasonable assumption for the model is that the rhyolitic cap, which is only part of the whole Rattlesnake Tuff magma chamber (Chapter II), has a flat cylindrical shape. Because of this cylindrical shape, it is assumed that most of the fractionation occurs along the top of the chamber which is also the coolest. Tank experiments simulating wall crystallization processes indicated that the roof can be the main crystallization site when cooling occurs primarily from roof downward and that compositional gradients through crystallization processes along the roof will only affect the uppermost layer (Baker & McBirney, 1985). The lower boundary of the rhyolite part of the chamber is an interface with dacite magmas (Chapter II). Mainly along the top of the chamber, the first more evolved rhyolite (group-D) is generated from the least-evolved rhyolite (group-E), which presumably was the rhyolite composition generated by partial melting (Fig. III-12a, b). Because the derivative daughter liquids is lighter is stays more or less where it was generated. The next fractionation step occurs when the first parent composition is largely shielded from the top of the chamber, the main fractionation site. At that point the first derivative liquid (composition D) becomes the next parent composition and the process repeats itself. If the same process happens several times then a compositionally zoned rhyolitic magma body, as recorded in the Rattlesnake Tuff rhyolites, can be generated by stacking down derivative liquids (Fig. III-12). Intra-layer convection would help preserve compositional layers. Is this scenario, changing gradients between rhyolite compositions, stepwise or continuous, would be largely controlled by changing fractionation processes along the walls involving crystal equilibria and possibly other fractionation mechanisms. Physical advantages of liquid fractionation processes occurring along the roof are: (1) derivative liquids stay where they are generated because they are lighter; therefore, there is no need for daughter liquids to move to a different storage site because a stable density stratification is generated from the

beginning, and (2) main differentiation site is roof which is likely to be the coolest and largest surface.

Conclusions

Chemical data of glassy pumices cluster in 4 to 5 groups suggesting several compositionally distinct high-silica magmas coexisted in the Rattlesnake Tuff magma chamber. Modal analysis on selected pumices shows that crystal content decreases from values around 1 weight % to aphyric in the most evolved composition. Progressive crystal chemical changes are evident for all major phases, namely feldspar, Fe-rich clinopyroxene, and magnetite and for the accessory phase zircon. Crystal chemical changes observed in the mafic phases and the dropping out of fayalite from the mineral assemblage is consistent with development of increasingly more oxidizing conditions with differentiation. Feldspar compositional data suggest crystallization under hypersolvus, water-undersaturated conditions and transition to subsolvus conditions in the most evolved high-silica rhyolites is suggested by sparse microphenocrysts in group A.

Partition coefficients for feldspars, clinopyroxenes, and magnetites of all mineral bearing high-silica rhyolite magmas change progressively with degree of differentiation for elements whose partitioning behavior correlate closely with crystal chemical changes. Other elements show limited ranges, excluding elements likely to be affected by trace phase inclusions.

Progressive partial melting, mixing, and crystal fractionation in modal proportions of observed phases are inadequate to explain chemical data. Assimilation is likely to have also played a minor role in the generation of the diversity of high-silica rhyolite magmas. Liquid fractionation along the margins in nonmodal mineral proportions seems the best way to generate several rhyolite magmas. It is proposed that the liquid fractionation process occurs primarily along the top of a presumably flat cylindrical shaped magma chamber. Along the roof of the magma chamber the next lighter and more evolved liquid is generated and stays as top layer. This way, each derivative liquid becomes the parent composition for the next more evolved rhyolite. Repetition of the process leads to a density and compositionally stratified high-silica rhyolite magma column.

Fig. III-13 Proposed differentiation model

Each more evolved liquid is generated from the previous less-evolved liquid mainly through derivative liquids obtained from wall fractionation processes along the top of the magma chamber. Stage (1) is stage where least evolved magma accumulated as coherent, homogenized magma batch; stage (6) represents stage of chemical zonation right before eruption. Arrows show schematically intra-layer convection and liquid-fractionation along the top and the sides of the magma chamber. A is most evolved, E is least evolved high-silica rhyolite composition comparative to observed composition of high-silica rhyolite pumice clusters, see text.











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